Modeling of future changes in seasonal snowpack and impacts on summer low flows in alpine catchments

Jenicek, Michal ; Seibert, Jan ; Staudinger, Maria

Abstract: It is expected that an increasing proportion of the precipitation will fall as rain in alpine catchments in the future. Consequently, snow storage is expected to decrease, which, together with changes in snowmelt rates and timing, might cause reductions in spring and summer low flows. The objectives of this study were (1) to simulate the effect of changing snow storage on low flows during the warm seasons and (2) to relate drought sensitivity to the simulated snow storage changes at different elevations. The Swiss Climate Change Scenarios 2011 data set was used to derive future changes in air temperature and precipitation. A typical bucket-type catchment model, HBV-light, was applied to 14 mountain catchments in Switzerland to simulate streamflow and snow in the reference period and three future periods. The largest relative decrease in annual maximum SWE was simulated for elevations below 2,200 m a.s.l. (60–75% for the period 2070–2099) and the snowmelt season shifted by up to 4 weeks earlier. The relative decrease in spring and summer minimum runoff that was caused by the relative decrease in maximum SWE (i.e., elasticity), reached 40–90% in most of catchments for the reference period and decreased for the future periods. This decreasing elasticity indicated that the effect of snow on summer low flows is reduced in the future. The fraction of snowmelt runoff in summer decreased by more than 50% at the highest elevations and almost disappeared at the lowest elevations. This might have large implications on water availability during the summer.

DOI: https://doi.org/10.1002/2017wr021648

Originally published at:
Jenicek, Michal; Seibert, Jan; Staudinger, Maria (2018). Modeling of future changes in seasonal snowpack and impacts on summer low flows in alpine catchments. Water Resources Research, 54(1):538-556.
DOI: https://doi.org/10.1002/2017wr021648

AGU PUBLICATIONS

Water Resources Research

RESEARCH ARTICLE

10.1002/2017WR021648

Key Points:
- In snow-dominated areas, summer low flows will significantly decrease in the future
- This decrease will be caused mostly by the decrease in snow storage and the shift of snowmelt season due to the increase in air temperature
- Low flows at higher elevations are more sensitive to the decrease in snow storage than low flows at lower elevations

Correspondence to:
M. Jenicek,
michal.jenicek@natur.cuni.cz

Citation:
Jenicek, M., Seibert, J., & Staudinger, M. (2018). Modeling of future changes in seasonal snowpack and impacts on summer low flows in alpine catchments. *Water Resources Research*, *54*, 538–556. https://doi.org/10.1002/2017WR021648

Received 9 AUG 2017
Accepted 3 JAN 2018
Accepted article online 9 JAN 2018
Published online 30 JAN 2018

Modeling of Future Changes in Seasonal Snowpack and Impacts on Summer Low Flows in Alpine Catchments

Michal Jenicek[1], Jan Seibert[2,3], and Maria Staudinger[2]

[1]Department of Physical Geography and Geoecology, Charles University, Prague, Czech Republic, [2]Department of Geography, University of Zurich, Zurich, Switzerland, [3]Department of Earth Sciences, Uppsala University, Uppsala, Sweden

Abstract It is expected that an increasing proportion of the precipitation will fall as rain in alpine catchments in the future. Consequently, snow storage is expected to decrease, which, together with changes in snowmelt rates and timing, might cause reductions in spring and summer low flows. The objectives of this study were (1) to simulate the effect of changing snow storage on low flows during the warm seasons and (2) to relate drought sensitivity to the simulated snow storage changes at different elevations. The Swiss Climate Change Scenarios 2011 data set was used to derive future changes in air temperature and precipitation. A typical bucket-type catchment model, HBV-light, was applied to 14 mountain catchments in Switzerland to simulate streamflow and snow in the reference period and three future periods. The largest relative decrease in annual maximum SWE was simulated for elevations below 2,200 m a.s.l. (60–75% for the period 2070–2099) and the snowmelt season shifted by up to 4 weeks earlier. The relative decrease in spring and summer minimum runoff that was caused by the relative decrease in maximum SWE (i.e., elasticity), reached 40–90% in most of catchments for the reference period and decreased for the future periods. This decreasing elasticity indicated that the effect of snow on summer low flows is reduced in the future. The fraction of snowmelt runoff in summer decreased by more than 50% at the highest elevations and almost disappeared at the lowest elevations. This might have large implications on water availability during the summer.

1. Introduction

Air temperature and precipitation are the most important drivers of snowpack variability at different elevations. With a future increase in air temperature more precipitation will fall as rain than as snow during the cold period (Foster et al., 2016; Harpold et al., 2017; Li et al., 2017; Zhang et al., 2015). These changes in snowfall fraction will lead to a decrease in snow accumulations. Marty et al. (2017) showed in two large Swiss catchments that snow storage might decrease by 50% at elevations above 3,000 m a.s.l. and almost no snow might accumulate at elevations lower than 1,200 m a.s.l. by the end of the twenty-first century. Similarly, the snowmelt runoff at elevations between 1,000 and 2,500 m a.s.l. will change more (in absolute values) due to future temperature changes than the snowmelt runoff at lower elevations (Speich et al., 2015).

Similar findings have been reported from the central Rocky Mountains, for which Sospedra-Alfonso et al. (2015) explored the importance of both air temperature and precipitation for snow accumulation at different elevations. They found a threshold elevation of 1,560 ± 120 m below which temperature is the main driver of the snowpack and above which precipitation is more important. A similar study has been performed by Morán-Tejeda et al. (2013) for Switzerland who found a threshold elevation at 1,400 ± 200 m. The increase in air temperature due to climate change also affects the snow cover duration. In general, the shorter total duration of the snow cover seems to be more related to earlier snowmelt rather than later snow onset (Klein et al., 2016). The combined effect of air temperature and precipitation was also investigated in the western United States using spatial and temporal analogs (Luce et al., 2014).

These studies indicate that elevation plays a key role for changes in snow storage and it fundamentally affects the sensitivity of catchments to water balance changes. Higher air temperatures during spring shift the onset of snowmelt and thus streamflow toward earlier spring (Barnett et al., 2005; Godsey et al., 2014; Langhammer et al., 2015). Additionally, the winter low flows, which are typical for mountain catchments, are expected to increase (Laaha et al., 2016). These changes lead to a higher streamflow occurring earlier in the water year (Blahušiaková & Matoušková, 2015; Hanel et al., 2012). On the contrary, earlier snowmelt also implies slower snowmelt rates due to the lower radiation in earlier spring compared to late spring and summer (Musselman et al., 2017).

Many studies showed that a decrease in snow affects groundwater recharge (Tague & Grant, 2004) and thus influences the streamflow during late spring and summer (Godsey et al., 2014; Jenicek et al., 2016). Berghuijs et al. (2014) showed that higher snowfall fraction leads to higher annual runoff in the contiguous United States. Additionally, with decreasing snowfall fraction and thus SWE, the spring and summer runoff will decrease (Barnhart et al., 2016; Brahney et al., 2017; Teutschbein et al., 2015).

The mentioned changes in snowfall fraction will affect not only the seasonal runoff volume, but also spring and summer low flows. The total amount of snow precipitation in winter affects groundwater recharge and hence also runoff during dry summer periods (Beaulieu et al., 2012; Van Loon et al., 2015). The period that is potentially affected by low flows toward late spring and early summer is expected to shift due to earlier snowmelt onset and melt-out (Etter et al., 2017). Snow that is accumulated in the cold period can affect low flows during the subsequent warm period especially in areas with large differences in winter and summer precipitation (Godsey et al., 2014). However, snow alone cannot explain the variability of low flow and drought, particularly in regions where precipitation is more equally distributed throughout the year (Jenicek et al., 2016). However, when looking at years with below-average liquid precipitation during spring and summer, snow affects low flows significantly more (Jenicek et al., 2016). In areas where precipitation has a pronounced seasonal character, as in the western United States, the role of winter precipitation and snow storage is much more important (Godsey et al., 2014).

These studies show that changes in snowpack and their influence on spring and summer runoff are widely studied and known. However, there is still limited information of how snowpack changes impact summer low flows across an elevation gradient and how this relation will change with predicted changes in air temperature and precipitation. Therefore, the main objectives of the study presented here were (1) to simulate the effect of changes in selected snow signatures on low flows during the warm season and (2) to relate drought sensitivity to the simulated snow storage. For this, we focused on snow changes and their impact on summer low flow at different elevations. The quantification of the changes in snow storage is important as there currently exist only few studies addressing this topic for the alpine region of central Europe. The consideration of different elevations is crucial as snow storage and its potential change due to climate change is highly variable with elevation.

2. Methods

2.1. Study Catchments

We selected 14 alpine and prealpine catchments in Switzerland with a catchment area ranging from 20 to 1,577 km^2 (Figure 1 and Table 1). Catchments were selected to be as close as possible to natural conditions, i.e., streamflow is near-natural and no major human influences such as dams or water transfers are present. Furthermore, the catchments are not at all or only to a minor degree covered by glaciers (area 0–2%) except for catchments C6 and C9 (up to 4%). A similar selection of study catchments has been used in previous studies focusing on future climate impacts assessment or low flows analyses (Addor et al., 2014; Jenicek et al., 2016; Staudinger et al., 2015; Staudinger & Seibert, 2014).

2.2. Data

Daily gridded precipitation and air temperature data (2 km resolution) were obtained from the Swiss Federal Office of Meteorology and Climatology (MeteoSwiss). These two data sets are based on daily observations of precipitation and air temperature measured at the high-resolution network in Switzerland (Frei, 2014; Frei & Schär, 1998). The data were averaged over the catchment area for the simulations and analyses. Daily snow water equivalent (SWE) data were also available as a gridded data set with a 1 km resolution. The SWE was calculated based on daily snow depth observations and a snow density model (Jonas et al., 2009) using interpolation and postprocessing procedures as first presented in Jörg-Hess et al. (2014). Daily time series were available for the period 1971–2012.

Daily streamflow data were provided by the Swiss Federal Office for the Environment (FOEN). Data for the period 1971–2012 were used for model calibration and validation except for two catchments for which data were available only since 1974 and 1975, as specified in Table 1.

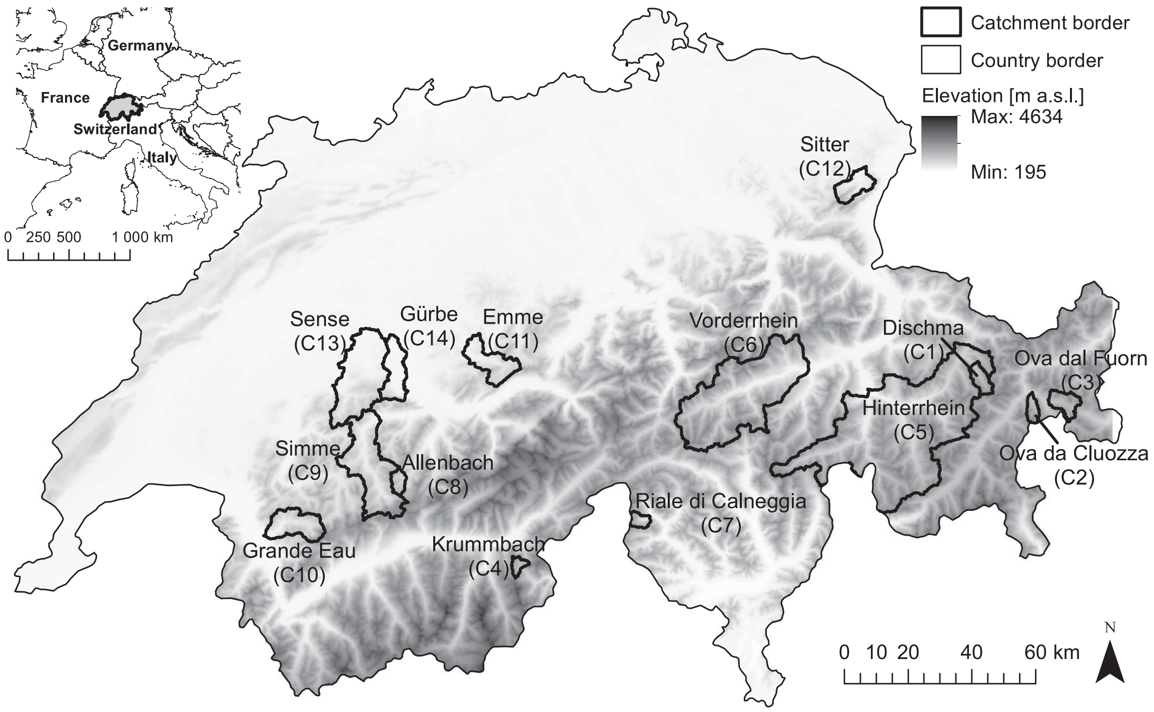

Figure 1. Location of the study catchments in Switzerland. Abbreviations from C1 to C14 are used consistently in the paper. The numbering indicates the order of the mean catchment elevation from the highest to the lowest.

2.3. Climate Scenarios

The Swiss Climate Change Scenarios 2011 data set (CH2011) was used to simulate the impact of future changes in air temperature and precipitation on catchment runoff (CH2011, 2011). The CH2011 data set provides daily estimates of changes in air temperature and precipitation relative to the reference period 1980–2009 for three future periods (2020–2049, 2045–2074, and 2070–2099, referred to as "scenario periods" in CH2011) using the A1B emission scenario. This data set contains the average daily difference of air

Table 1
Study Catchments and Selected Characteristics

Abbreviation (catchment name, gauging station)	Area (km^2)	Mean elevation (m a.s.l.)	Elevation range (m a.s.l.)	Mean slope (°)	Mean SWE_{max} (mm)	Snowfall fraction (–)	Observed data from (to 2012)	Climate region
C1 (Dischmabach, Davos)	42.9	2,368	1,667–3,138	22.9	484	0.97	1971	Northeast
C2 (Ova Da Cluozza. Zernez)	27.0	2,361	1,507–3,160	26.8	339	0.98	1971	Northeast
C3 (Ova Dal Fuorn, Zernez)	55.3	2,328	1,706–3,156	18.9	339	0.97	1971	Northeast
C4 (Krummbach, Klusmatten)	19.8	2,276	1,795–3,269	18.3	474	0.85	1971	South
C5 (Hinterrhein, Fürstenau)	1,577	2,113	649–3,406	21.9	351	0.91	1974	Northeast
C6 (Vorderrhein, Ilanz)	774	2,023	691–3,605	23.0	442	0.88	1971	Northeast
C7 (Riale di Calneggia, Cavergno)	23.9	1,986	883–2,911	29.1	423	0.88	1971	South
C8 (Allenbach, Adelboden)	28.8	1,851	1,296–2,753	19.7	351	0.78	1971	West
C9 (Simme, Oberwil)	344	1,632	776–3,242	18.1	264	0.74	1971	West
C10 (Grande Eau, Aigle)	132	1,557	417–3,204	21.1	249	0.71	1971	West
C11 (Emme, Eggiwil)	124	1,275	581–2,220	14.2	185	0.59	1975	West
C12 (Sitter, Appenzell)	74.4	1,247	769–2,501	17.8	193	0.62	1971	Northeast
C13 (Sense, Thörishaus)	351	1,068	551–2,181	9.9	94	0.39	1971	West
C14 (Gürbe, Belp)	116	845	518–2,169	8.7	51	0.41	1971	West

Note. Mean SWE_{max} is the mean annual SWE maximum of the observed period; snowfall fraction is the rate of snowfall to total precipitation of the period from November to April (see section 2.5 for detailed explanation). Climate region column indicates the region used in climate change scenarios according to the CH2011 data set (*CH2011*, 2011).

10.1002/2017WR021648

temperature between the reference period and given future period and respective scaling factor for the precipitation resulting from 10 different regional climate models (RCM). Daily values in this data set were derived from seasonal means using the concept of harmonic components (CH2011, 2011). The CH2011 data set provides daily values for three Swiss regions; northeast, west, and south. The individual catchments were divided according to these three regions as indicated in Table 1.

2.4. HBV Model

2.4.1. Model Structure, Calibration, and Validation

To quantify the impact of predicted changes in air temperature and precipitation on snow storage and streamflow several model experiments were set up using a typical bucket-type catchment model. We here used the HBV model (Hydrologiska Byråns Vattenbalansavdelning; Lindström et al., 1997) in its software implementation HBV-light (Seibert & Vis, 2012). The HBV model consists of four basic routines to simulate catchment runoff:

1. The snow routine uses a degree-day approach, which calculates the snow accumulation and snowmelt including snow water holding capacity and potential refreezing of meltwater.
2. The soil routine includes groundwater recharge and actual evaporation, which are simulated as functions of the actual water storage in the soil box.
3. The response routine calculates the catchment runoff as a function of water storage in an upper and a lower groundwater box.
4. The routing routine uses a triangular weighting function propagating the runoff to the outlet of the catchment.

The main inputs form time series of daily precipitation, daily air temperature, and monthly potential evapotranspiration (PET). The monthly PET values were calculated using simple temperature-based method presented in Oudin et al. (2005). The model uses a linear interpolation to calculate daily values. Observed streamflow is used for model calibration and validation. In our experimental setup, the model additionally needs observed daily SWE as this variable is used for model calibration in addition to streamflow. The study catchments were subdivided into elevation zones of 100–200 m, which was a compromise between reflecting the variations of precipitation and temperature with elevation and an unnecessary computational demand.

The HBV model was calibrated for each catchment with a genetic calibration algorithm by which optimized parameter sets were found by consecutive evolution of parameter sets using selection and recombination (Seibert, 2000). The integrated multivariable model calibration procedure was used to calibrate the model (Rientjes et al., 2013; Seibert, 2000). A combination of four criteria served to evaluate the goodness of fit of the model (Table 2): (1) model efficiency for runoff (R_{runoff}) (Nash & Sutcliffe, 1970), (2) model efficiency for SWE (R_{SWE}), (3) volume error (R_{vol}), and (4) mean absolute relative error (R_{MARE}). The objective function R_{weighted} was a combination of these four criteria giving the different weights a, b, c, and d for each criterion. Different weights were tested for each catchment individually to achieve the best possible values of R_{weighted} during calibration of both runoff and SWE (Table 2). The aim was to reproduce the current catchment behavior considering different aspects of the streamflow response as well as snow accumulation and melt in order to get reliable future simulations.

Table 2
Objective Functions Used for Model Calibration and Validation

Objective function	Equation	Weights (tested separately for each catchment)
Model efficiency for runoff	$R_{runoff}=1-\frac{\sum(Q_{obs}-Q_{sim})^2}{\sum(Q_{obs}-\bar{Q}_{obs})^2}$	20%
Model efficiency for SWE (S)	$R_{SWE}=1-\frac{\sum(S_{obs}-S_{sim})^2}{\sum(S_{obs}-\bar{S}_{obs})^2}$	20–40%
Volume error	$R_{vol}=1-\frac{\lvert\sum(Q_{obs}-Q_{sim})\rvert}{\sum(Q_{obs})}$	20–30%
MARE measure	$R_{MARE}=1-\frac{1}{n}\sum\frac{\lvert Q_{obs}-Q_{sim}\rvert}{Q_{obs}}$	20–40%
Weighted efficiency	$R_{weighted}=a\cdot R_{runoff}+b\cdot R_{SWE}+c\cdot R_{vol}+d\cdot R_{MARE}$	n.a.

A split sample test was used for model calibration and validation. The model was calibrated using data from 1971 to 1991 (except two catchments, see Table 1). The model was validated using data from 1992 to 2012. Snow conditions between these two periods particularly changed especially for mean maximum SWE which decreased by 23%. However, the interannual variability did not change significantly.

Different calibration trials might result in different parameter sets with similar model performances during calibration but different behavior during other periods. To better address this parameter uncertainty, the model was calibrated 100 times resulting in 100 “best” parameter sets. These 100 sets were then used to simulate 100 different time series. Most of the further analyses were then based on these 100 series for both reference and future periods.

2.4.2. Simulation of the Reference and Future Periods

We performed hypothetical simulations, which allowed us to analyze the effect of snow storage on low flows separately from other water balance components (mainly liquid precipitation in the warm season and actual evapotranspiration, AET).

The reference period 1980–2009 was created as a simulation of previous catchment behavior (using observed input data and calibrated parameter sets). The climate scenario periods were constructed in two consecutive steps. First, the daily air temperature and precipitation changes from the CH2011 data set were applied (additive changes for air temperature and multiplicative changes for precipitation) to the respective time series for the reference period 1980–2009 to create three future periods. Second, the resulting time series of air temperature and precipitation were used for simulation using the parameter sets of the model calibration. This procedure does not reflect a possible change of the interannual variability in air temperature and precipitation in the future.

In the climate scenarios, PET was kept constant despite the increase in air temperature. That enabled us to better separate the effect of increased air temperatures on snowfall fraction and snow storages and thus changes in minimum runoff during the warm period. In this way, we could attribute simulated changes in summer minimum runoff to changes in winter conditions and to changes in spring and summer liquid precipitation.

The HBV model simulates daily time series of several variables such as SWE, AET as well as total runoff and runoff originating from snowmelt. The simulated values of the variables were calculated both as catchment means and as mean values for each elevation zone. For further analysis, both the simulation mean for the catchment and simulation means per elevation zone were used (for each catchment and each of the 100 parameter sets). The HBV model outputs are publicly available (Jenicek et al., 2017).

2.4.3. Changes in Elasticity of Summer Low Flows to Changes in SWE

Changes in spring and summer streamflow due to changes in snowpack are driven by both the total amount of snow per season and the timing of snowmelt. We set up two modeling experiments to partly separate the total amount of snow and the timing of snowmelt. Additionally, we were interested in how the sensitivity of different catchments would change with changes in either total amount of snow or timing of snowmelt. The snow routine in the HBV model contains two parameters governing snow storage and snowmelt timing. One parameter is the threshold temperature T_T that differentiates between snow and rain and sets the air temperature of snowmelt onset. The other parameter is the snowfall correction factor S_{FCF} that accounts for snow undercatch due to wind. S_{FCF} is a relative value used to adjust precipitation whenever it is simulated as snow. With the change of these two parameters we can quantify to which degree changes in low flows are caused by a decrease in SWE or by an earlier snowmelt onset.

Both mentioned modeling experiments have been applied only for the reference period. In the first modeling experiment, we progressively changed the threshold temperature T_T from $-5°C$ to $+5°C$ (in steps of 0.1°C). In the second modeling experiment, the same have been done with introducing a progressive change in S_{FCF} from 0.1 to 2.0 (in steps of 0.1). Changes in both parameters influenced the simulated snowfall and thus SWE. Therefore, melt-out was influenced as well. However, T_T additionally influences the timing of snowmelt onset and snowmelts rates. Thus, changing T_T controls both snow amount and melt-out in a similar way, while changing S_{FCF} controls mainly the amount of snow while the timing of melt-out is affected less.

2.5. Snow and Streamflow Signatures Used to Analyze Snowpack Changes and Summer Low Flows

We selected four snow and three streamflow signatures to analyze the effect of changes in snow storage on low flows (Table 3). These characteristics were calculated separately for each parameter set (100 sets) and simulation period. First, 100 different values of the same signature were calculated for each catchment and simulation period. Then, the median from these 100 values was calculated and used for further analysis. This approach increased the robustness of the individual signatures resulting from different model parameterizations.

2.5.1. Snow Signatures

The effect of the changed climate series and model settings on snow was quantified using different snow signatures. Annual maximum SWE represents late winter conditions using the simulated SWE data from February to May (SWE_{max}) to avoid possible early winter snow peaks. In a few rare cases, SWE_{max} in lower catchments did not represent the maximum annual SWE, but the maximum already occurred before February.

Additionally, the day of the year with maximum SWE (*DOY of* SWE_{max}) was extracted to describe the start of the snowmelt season.

Table 3
Snow and Streamflow Signatures

Snow signatures
Maximum of SWE from February to May (including) before melting calculated both for elevation zones and catchment mean (SWE_{max})
Day of year with maximum SWE (*DOY of* SWE_{max})
Melt-out date (*DOY of melt-out*)
Snowfall fraction (rate of snowfall to total precipitation from November to April)
Streamflow signatures
Minimum of 7 day moving average of runoff calculated separately for each month from March to August or for season from June to August (Q_{min})
Flow duration curve calculated from June to August (FDC)
Fraction of runoff from snow to total runoff calculated from June to August (Q_{sf})

The end of winter was defined using the melt-out date (*DOY of melt-out*). *DOY of melt-out* dates were extracted from simulated SWE data for each catchment and year. We defined the melt-out as the first occurrence of SWE below a threshold of 20 mm after the *DOY of* SWE_{max}. We also tested other threshold values and found that this did not influence the results considerably.

The phase of the precipitation in the cold period was described by the snowfall fraction. The snowfall fraction is the fraction of snowfall water equivalent (hereafter referred to as snowfall) to total precipitation calculated for the period from November to April. To distinguish between snow and rain, a single threshold temperature, model parameter T_T, was used. The value of T_T was calibrated individually for each catchment.

2.5.2. Streamflow Signatures

The summer low flows were described using the minimum 7 day moving average of daily runoff (mm d^{-1}), which was calculated from the HBV simulations (hereafter referred as minimum runoff, Q_{min}). We also tested other widths of the moving window (3 and 15 days) but there was no significant influence on the results. The monthly minimum runoff (March–August) or the seasonal minimum runoff (June–August) was used in all analyses.

A more comprehensive measure for low flow than a single value is the flow duration curve (FDC). FDCs were calculated for all study catchments for the reference period and all future periods. The FDC of the specific catchment shows the exceedance probability of mean daily runoff. Only daily runoff from June to August was selected to calculate the FDCs since this study is focused on changes in minimum runoff in spring (after melt-out) and summer.

The proportion of the different runoff components is an additional information to quantify the origin of streamflow. The HBV model keeps track of whether the runoff originates from snow (Q_s) or rain (Q_r) throughout the model routines. The procedure to track both rain and snow components is based on complete mixing in all reservoirs. There are two assumptions used during calculation, which are necessary to mention (1) the liquid water content of the snowpack is either considered as rain (in case of rain on snow) or snow (in case of melt of the snowpack) and (2) the only time that the source of the water can change is when refreezing in the snowpack is simulated. As soon as refreezing occurs, the source is considered as snow. The fraction of runoff from snow to total runoff calculated from June to August (Q_{sf}) was used as a streamflow signature in the analysis.

Statistical regressions describe the relationship between the relative change in monthly minimum runoff and the relative change in maximum SWE. We used annual SWE_{max} and monthly Q_{min} for all years in the reference and future periods. The nonparametric Theil-Sen slope of the regression was used to evaluate our statistical models. The Theil-Sen slope is a median of slopes calculated for each pair of observations (Wilcox, 2001). The Theil-Sen regression model is suitable for nonnormally distributed data with outliers. The elasticity index E was then calculated by dividing the relative change in monthly minimum runoff (dQ_{min}) by the relative change in maximum SWE, dS_{max}

$$E = dQ_{min}/dS_{max} \tag{1}$$

Each catchment was analyzed separately using annual SWE_{max} and monthly or seasonal Q_{min} for the reference period and the three future periods. This way the change of importance of snow contribution to low flows was highlighted in different catchments and at different times.

3. Results

3.1. Model Calibration and Validation

The values of individual objective functions for both calibration and validation periods indicate that the use of the combined objective function resulted in good overall model performances (Figure 2). Model simulated correctly both runoff (relatively high values of R_{runoff}, R_{vol}, and R_{MARE}) and SWE (R_{SWE}). The reliable

10.1002/2017WR021648

simulation of the catchment snow storage was important especially for higher elevation catchments with high snow storage.

3.2. Changes in Meteorology, Snow, and Streamflow Signatures

As expected, there was a strong decrease in signatures describing snow storage for the three future periods compared to the reference period for all study catchments and elevations (Figure 3). The largest absolute decrease in maximum annual SWE is predicted for elevations from 2,000 to 2,700 m a.s.l. (Figure 3, top left). A relative decrease by 30–50% for the first future period is predicted for elevations below 2,200 m a.s.l. and this decrease is even stronger for the future period 2070–2099 with up to 80% compared to the reference period. Above 2,200 m a.s.l., the relative decrease in SWE_{max} for the third future period is lower with up to 60% for elevations between 2,200 and 2,500 m a.s.l. and 20–40% for elevations higher than 2,500 m a.s.l. Looking at the snow routine of the HBV model, it is clear that the decrease in simulated SWE_{max} was mostly caused by the decrease in snowfall fraction (by 40% at elevations around 2,000 m a.s.l., Figure 3, top right) due to the increase in air temperature (Figure 4). The future changes in precipitation in the cold period will be rather minor (Figure 4). Therefore, their influence on changes in SWE can be assumed to be minor as well. The air temperature increase resulted both in earlier *DOY of* SWE_{max} and *DOY of melt-out*. Additionally, the snowmelt period became shorter (Figure 3, bottom plots) especially for elevations above 1,500 m a.s.l. (*DOY of melt-out* was shifted by 40–50 days toward earlier spring, *DOY with* SWE_{max} was shifted only by 30–40 days).

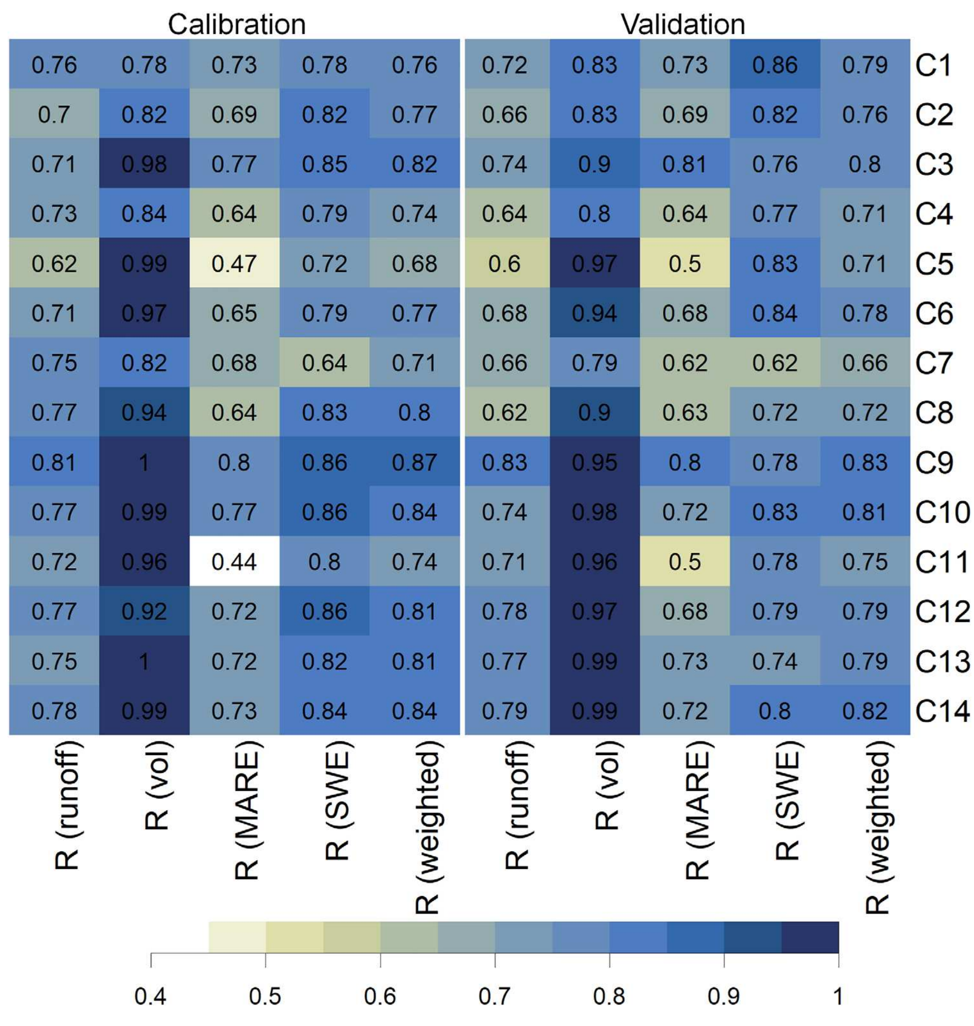

Figure 2. Values of individual objective functions for each catchment for calibration and validation periods (mean from 100 parameter sets). Color scale shows values of individual calibration criteria.

To make a link between changes in snow signatures and other climate drivers, we analyzed the changes in seasonal air temperature, precipitation, and simulated AET (Figure 4). The climate change scenarios indicated a strong increase in mean air temperature by 3–4°C in the warm period (June–August) and by 2–3°C in the cold period (November–April) in all elevation zones (Figure 4, left plots). The predicted precipitation in the warm period decreases on average by 21%, but increased in the cold period by 3–5% (Figure 4, middle plots). The changes in air temperature and precipitation in the warm period will affect actual evapotranspiration (AET) in the warm period, which will increase by 5% in the first future period compared to the reference period and slightly decrease in the second and third future periods in the catchments higher than 2,000 m a.s.l. However, AET is projected to decrease by 12% in lower elevation catchments (<1,500 m a.s.l.) for the future period 2070–2099 compared to the reference period (Figure 4, top right plot). This decrease in AET may be caused by reduced water availability. On the contrary, the gradual extension of the period without snow cover temporarily increases the AET because in the HBV model AET is assumed to be zero

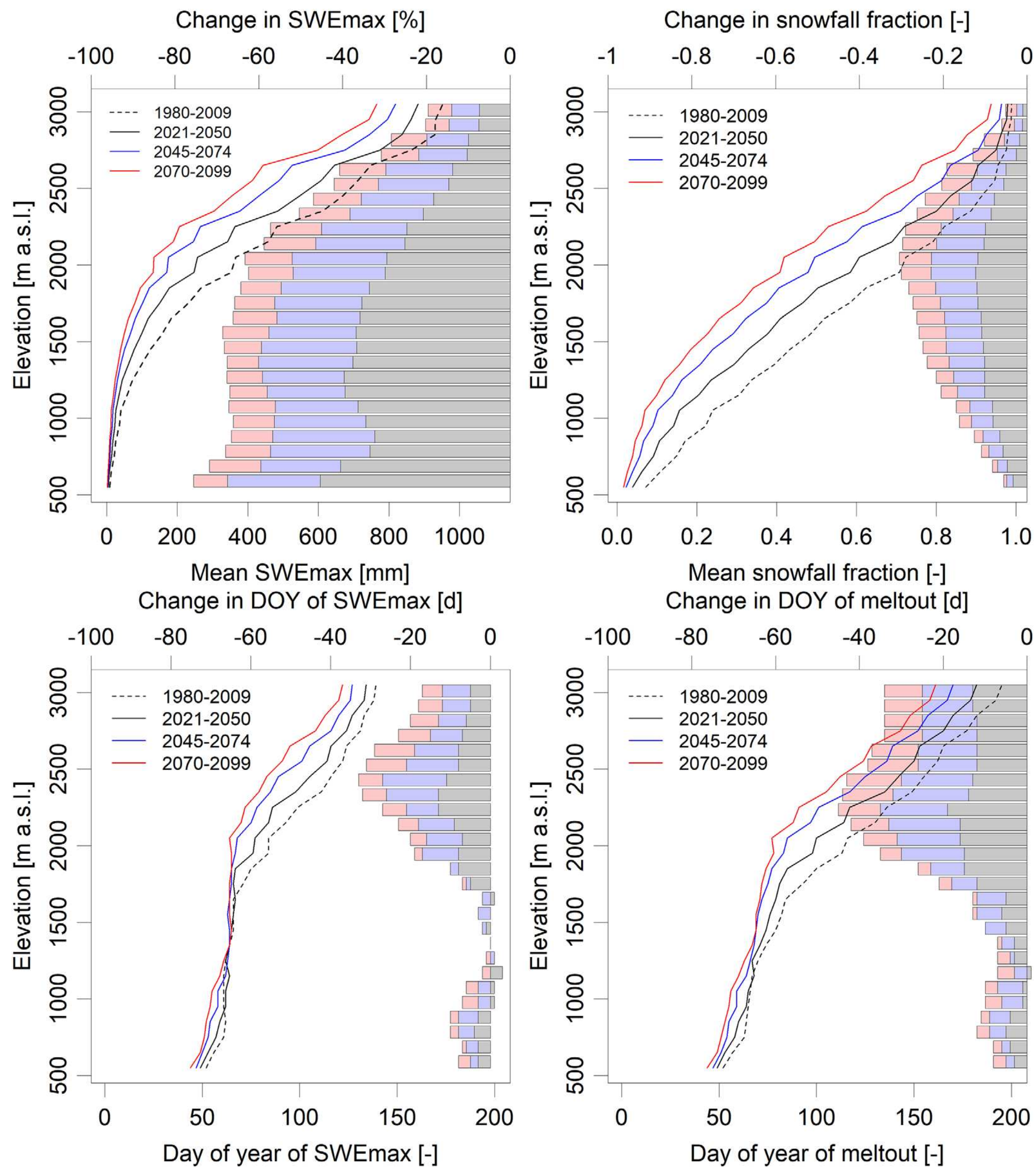

Figure 3. (top left) Mean SWE_{max}, (top right) mean snowfall fraction, (bottom left) DOY of SWE_{max}, and (bottom right) DOY of melt-out at different elevations for the reference period and three future periods. Lines express real values, bars express relative differences from the reference period.

when snow cover occurs. The mentioned fact resulted in strong relative increase in simulated AET in the cold period (by 50–100% with latter for highest elevations), although the absolute increase in AET is not so large (20–50 mm depending on elevation and future period, Figure 4, bottom right plot). It is important to mention that our modeling experiments did not account for PET changes (see section 2) thus the AET changes may be explained only by changes in precipitation, snow cover duration and overall water availability in the soil box.

The changes in summer streamflow were described using flow duration curves (FDCs). The FDCs were calculated separately for each study catchment for the reference period and the three future periods (Figure 5). The derived FDCs showed a change of exceedance probability for the period from June to August with large runoff decrease during this period. This decrease is stronger for catchments above 1,500 m a.s.l. compared to catchments at lower elevations. Therefore, simulated changes may be attributed mostly to the combined effect of snowmelt season shift toward earlier spring, decrease in SWE and decrease in precipitation. The decrease in runoff volume is evident mostly for exceedance probabilities between 30% and 70% but it is also clearly visible for runoff minima (decrease in Q_{90} by 32–56% in the future period 2070–2099 compared to the reference period).

The flow duration curves calculated for the whole year (not shown in the paper) indicated that there is only a small decrease in total runoff volume in future periods compared to the reference period. The decrease in runoff volume during the warm period was partly compensated by the increase in runoff volume during the cold period. The decrease in total runoff volume may be related mostly to the decrease in precipitation in the warm period. Although, AET in the warm period decreased as well, this decrease can only partly compensate the effect of decrease in precipitation.

3.3. Influence of Snow Storage on Spring and Summer Low Flows

The relation between annual maximum snow storage (SWE_{max}) and monthly minimum runoff (Q_{min}) has been analyzed at a catchment scale separately for each month from March to August. The elasticity was

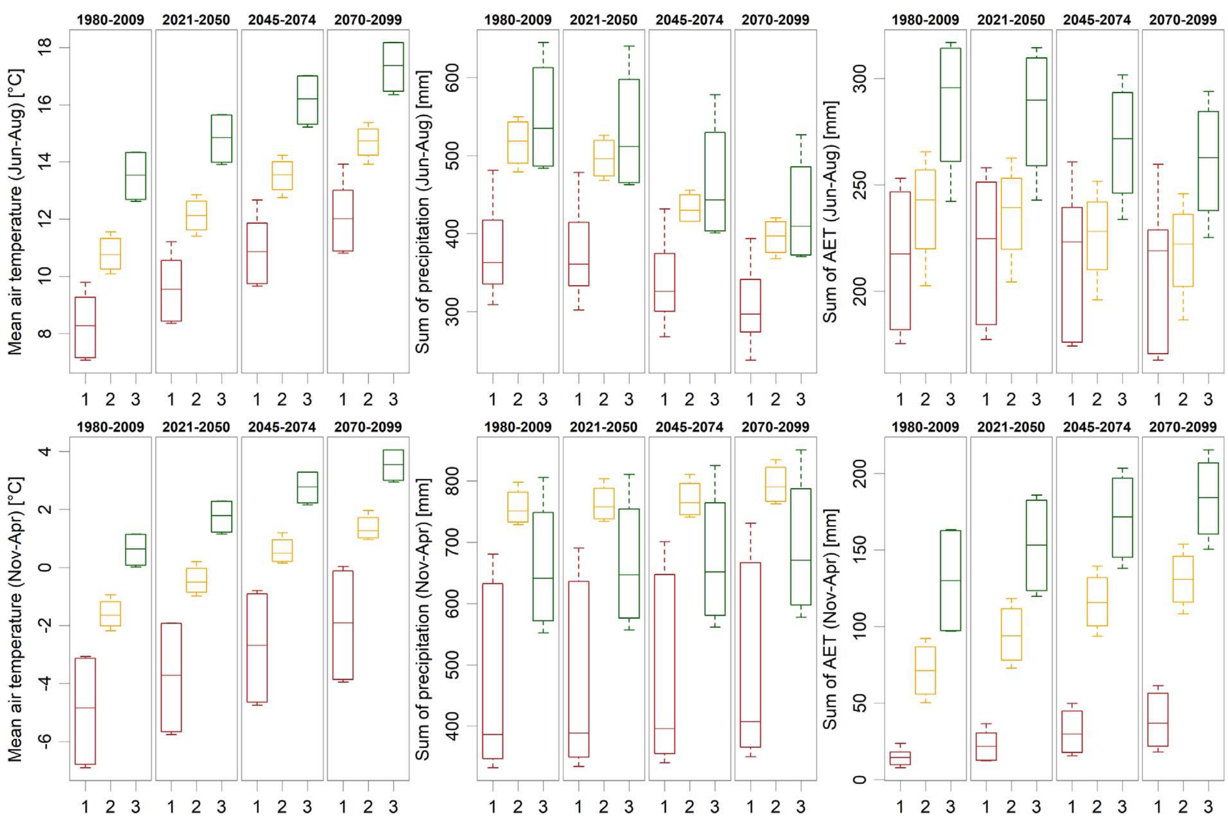

Figure 4. Changes in mean air temperature, mean precipitation, and mean AET from June to August and from November to April simulated by the HBV model for the reference period and the three future periods. Group 1 (brown): catchments with mean elevation >2,000 m a.s.l.; group 2 (dark yellow): catchments between 1,500 and 2,000 m a.s.l.; and group 3 (green): catchments <1,500 m a.s.l.

10.1002/2017WR021648

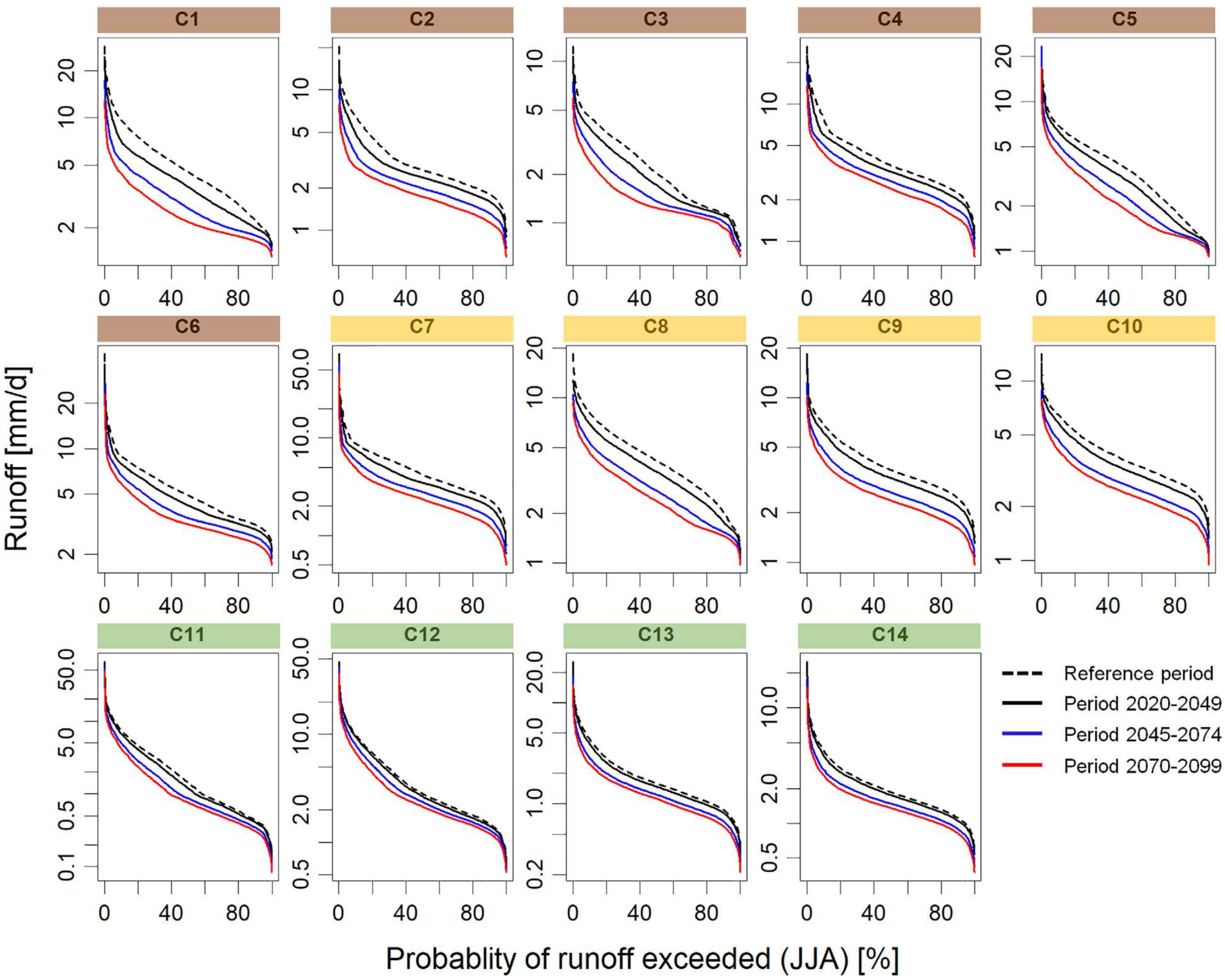

Figure 5. Flow duration curves for individual study catchments. The curves show the probability of runoff exceedance from June to August for the reference and future periods. Background color for catchment name represents catchment elevation group. Group 1 (brown): catchments with mean elevation >2,000 m a.s.l.; group 2 (dark yellow): catchments between 1,500 and 2,000 m a.s.l.; and group 3 (green): catchments <1,500 m a.s.l. Note that the *y* axis has a logarithmic scale.

calculated from Theil-Sen's slopes for each catchment and month for the reference period and all three future periods. Only statistically significant relations are shown in Figure 6 ($\alpha = 0.1$). The annual SWE_{max} was correlated to monthly Q_{min} from May to July or August in catchments with a mean elevation higher than 2,000 m a.s.l. for the reference period (catchments C1–C6, except catchment C4). Here a relative decrease in SWE_{max} by 10% caused a relative decrease in monthly Q_{min} by 4–9% (see color scale in Figure 6). For some higher elevation catchments (C1, C4, and C5) in May, and for most of catchments in March and April, the elasticity was lower or not statistically significant. These lower elasticities were caused by remaining snow accumulation conditions rather than snowmelt in these months, which means that the relation between snow storage and runoff is small or not significant. In most of catchments higher than 1,500 m a.s.l. (C1–C9), the snow usually remained until June (see black points in Figure 6).

The monthly minimum runoff responded to maximum snow storage only from April to June for the future period 2070–2099 (Figure 6). Here the elasticity reached 0.3 to 0.9 in April and May, but decreased in June. The correlation between SWE_{max} and monthly minimum runoff was statistically not significant for July and August. For catchments with mean elevation less than 1,800 m a.s.l. (C8–C14), the correlation between SWE_{max} and monthly Q_{min} was significant for May and June for the reference period and progressively decreased across all future periods.

A significant decrease in seasonal minimum runoff from June to August was simulated for all future periods compared to the reference period (Figure 7). Full-colored marks indicate years when annual SWE_{max} for an

10.1002/2017WR021648

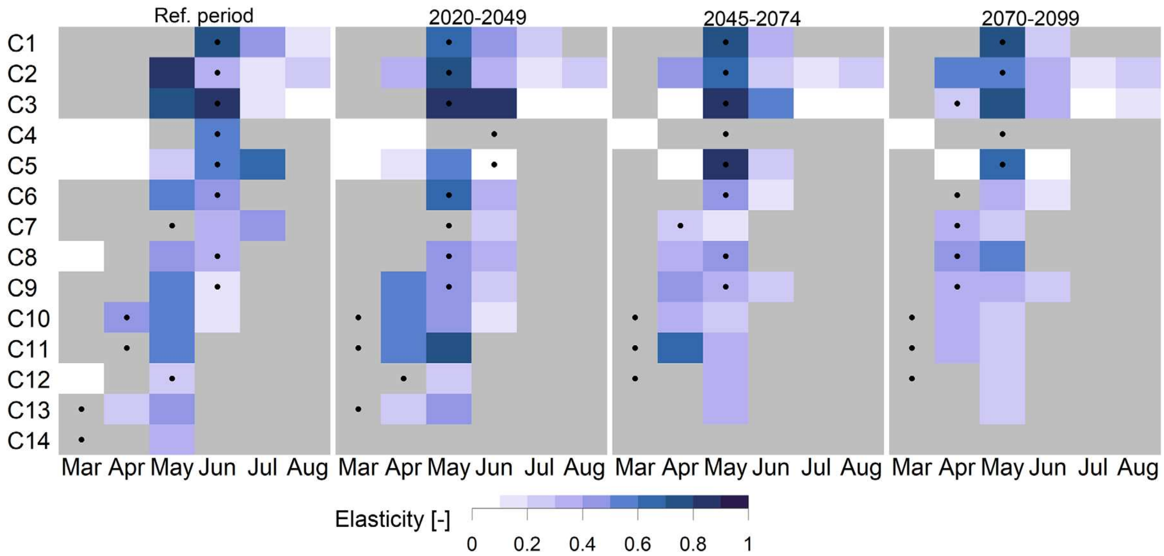

Figure 6. Dependence of monthly Q_{min} on annual SWE_{max} for all studied catchments (sorted by elevation from highest (C1) to lowest (C14)) for the period from March to August and for the reference and future periods. The colors present the values of Theil-Sen's slope and thus elasticity. Grey color used for relations that were not statistically significant at the 0.1 level. Black points indicate an average month of snow melt-out in individual catchment for all simulation periods.

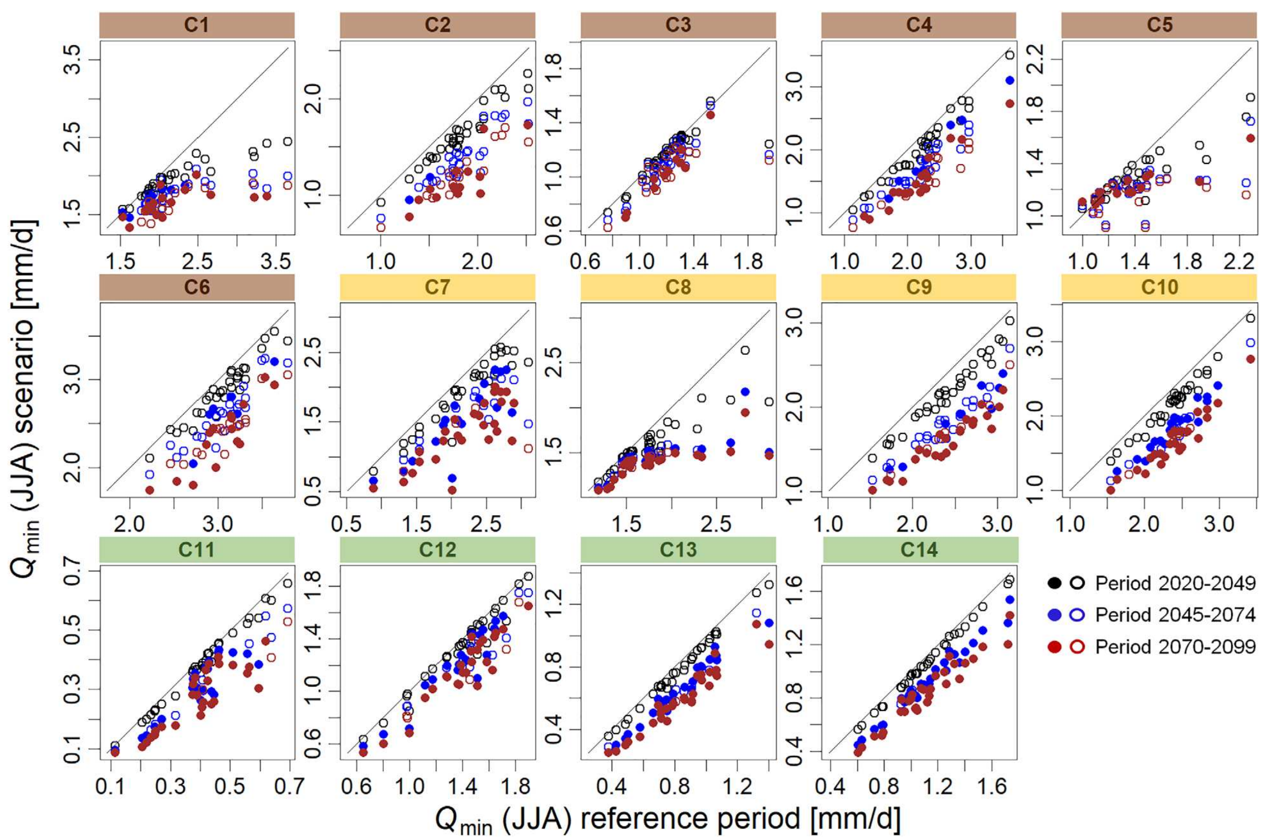

Figure 7. Seasonal minimum runoff (Q_{min}) from June to August (JJA) for the three future periods compared to the reference period. Full-colored marks indicate years when annual SWE_{max} for an individual future period decreased to less than 50% of the respective annual SWE_{max} in the reference period. Background color for catchment name represents catchment elevation group. Group 1 (brown): catchments with mean elevation >2,000 m a.s.l.; group 2 (dark yellow): catchments between 1,500 and 2,000 m a.s.l.; and group 3 (green): catchments <1,500 m a.s.l. Note different scaling of axes in individual plots.

individual future period decreased to less than 50% of the respective annual SWE_{max} in the reference period. The results for individual catchments (shown in individual plots) clearly indicated that the decrease in seasonal minimum runoff from June to August may be partly related to the decrease in maximum annual SWE. The relative decrease in seasonal minimum runoff is similar for both years with low minimum runoff and years with high minimum runoff. However, for some higher elevation catchments (C1, C5, C7, and C8), it seems that the relative decrease in seasonal minimum runoff in future periods compared to the reference period is bigger for years with higher seasonal minimum runoff. This might indicate that higher values of minimum runoff are more influenced by snow storages in these catchments and thus more sensitive to the decrease in SWE_{max}.

Since the scenario simulations do not reflect the changes in PET, decrease in low flows in spring and summer may be attributed partly to the decrease in snow storages and partly to the decrease in summer precipitation. Additionally, the importance of snow storages to explain the decrease in seasonal minimum runoff from June to August may be supported by the fact that the fraction of runoff volume originating from snowmelt to total runoff from June to August (Q_{sf}) strongly decreased for all future periods compared to the reference period (Figure 8). This decrease in runoff from snow may correlate to the decrease in total minimum runoff. The decrease in minimum runoff from June to August is largest in most of higher elevation catchments (C1–C10) which are characterized by high snow storages and by the largest absolute decrease in snow storage for future periods compared to the reference period.

In most catchments, the change of summer low flow is smaller than the change of runoff from snow and this is more obvious in catchments at lower elevations (C11–C14; Figure 8). This indicates that besides snow, there are still other factors, which influence the decrease in low flow, such as the decrease in summer

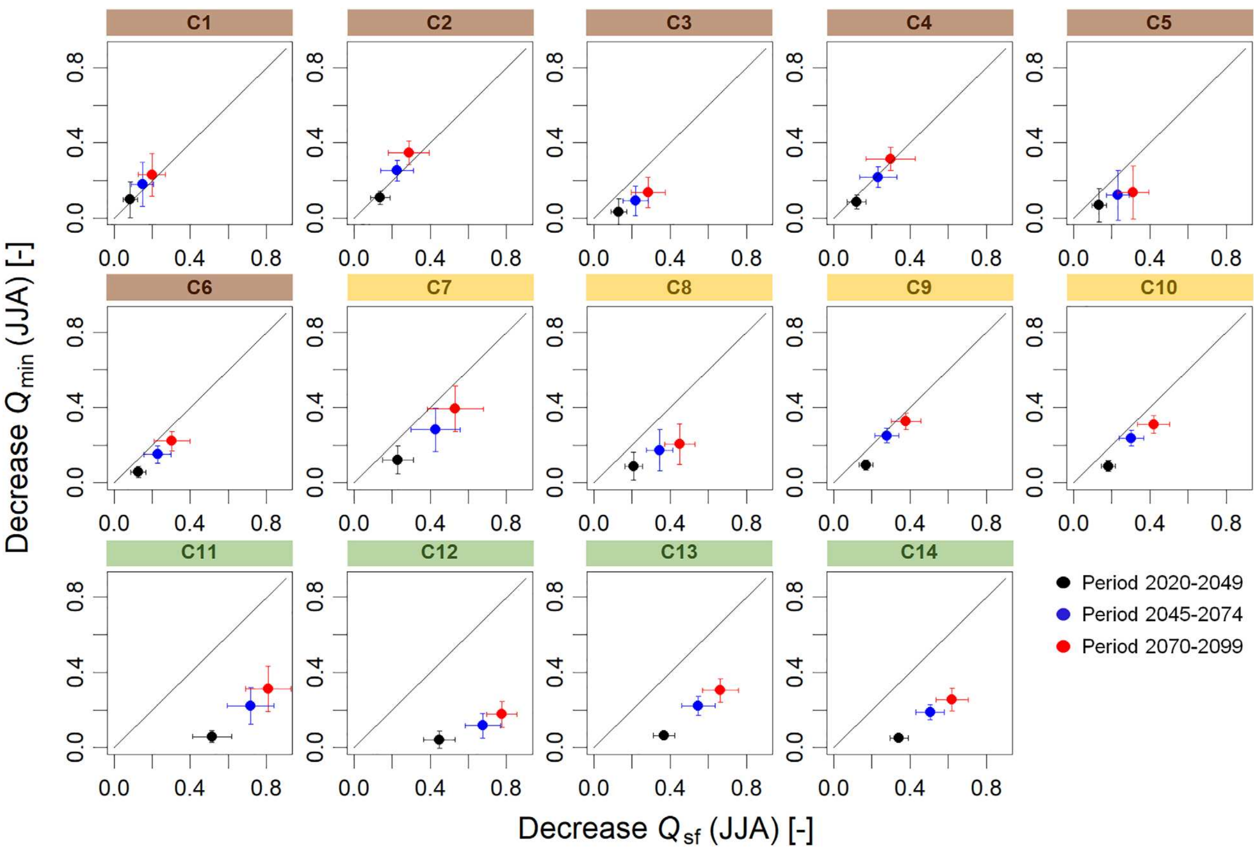

Figure 8. Decrease in Q_{sf} compared to the decrease in Q_{min} from June to August (JJA) for future periods compared to the reference period. Marks indicate mean value and error bars indicate the interannual variability (standard deviation) within individual future periods. Background color for catchment name represents catchment elevation group. Group 1 (brown): catchments with mean elevation >2,000 m a.s.l.; group 2 (dark yellow): catchments between 1,500 and 2,000 m a.s.l.; and group 3 (green): catchments <1,500 m a.s.l.

precipitation and increase in AET. The importance of these two factors increases in lower elevation catchments compared to higher elevation catchments.

3.4. Changes in elasticity of Summer Low Flows With Respect to Changes in SWE

To describe the changes in elasticity in different snow conditions and to investigate the combined effect of SWE decrease and the shift of snowmelt season, we performed two modeling experiments applied to the reference period time series; first with changing threshold temperature T_T, second with changing snowfall correction factor S_{FCF} (see section 2). We then analyzed the simulated changes in annual SWE_{max} and seasonal Q_{min} (JJA), specifically how summer minimum runoff changed due to changes in maximum SWE (i.e., elasticity). The elasticity was derived from a regression slope between two neighboring values of SWE_{max} and related to seasonal Q_{min} (Figure 9).

The elasticity clearly changes with maximum SWE. The elasticity is always positive, which means that there is always positive correlation between SWE_{max} and Q_{min}. The elasticity did not change with changes in SWE_{max} for the highest two catchments (C1 and C2), but it changed for the rest of catchments. In general, when SWE_{max} decreases, the elasticity decreases as well (there is too little snow to influence the runoff), but starting from a certain point, the elasticity also decreases with increasing SWE_{max}. The latter is more obvious for the experiment with changing T_T (red line) which combined the influence of both changes in snow storage and melt-out compared to the experiment with changing S_{FCF} for which the effect of changes in snow storage was crucial (and changes in melt-out happen only because of higher snow storages). The decrease in elasticity can be explained by large snow storage that does not melt during the warm season. Thus, its influence on summer runoff decreases. In general, it seems that middle and higher elevation catchments showed higher sensitivity of low flows to changes in both T_T and S_{FCF} than lower elevation catchments (Figure 9). The results indicated that the decrease in SWE affected low flows slightly more than earlier snowmelt (on average, the sensitivities for T_T is higher than for S_{FCF}).

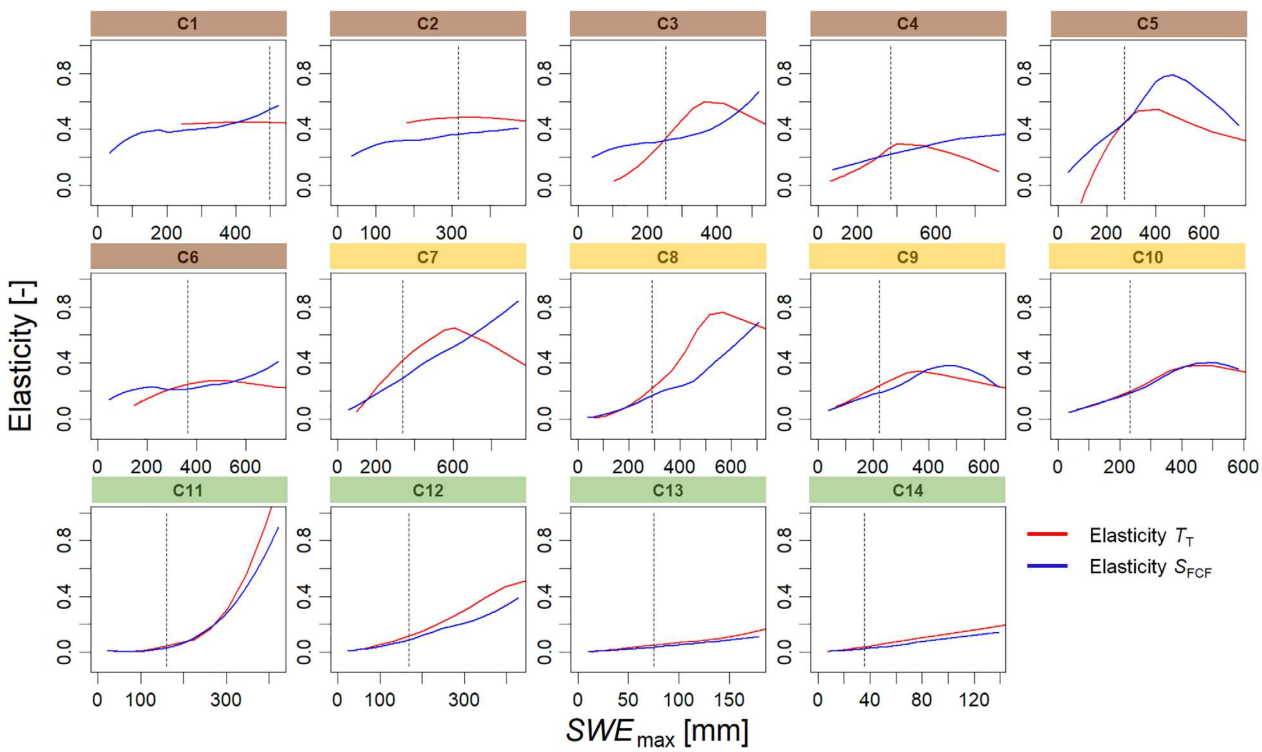

Figure 9. Relation between elasticity and SWE_{max} for study catchments for the experiment with changing threshold temperature T_T (red line) and for the experiment with changing snowfall correction factor S_{FCF} (blue line). Vertical dashed line is mean SWE_{max} in the catchment for the reference period. Background color for catchment name represents catchment elevation group. Group 1 (brown): catchments with mean elevation >2,000 m a.s.l.; group 2 (dark yellow): catchments between 1,500 and 2,000 m a.s.l.; and group 3 (green): catchments <1,500 m a.s.l.

4. Discussion

4.1. HBV Model Calibration and Validation

The presented results are based on modeling experiments assuming future increase in air temperature and changes in precipitation. The uncertainty arising from the model parameterization was addressed by the ensemble approach using 100 parameter sets, which generated more robust results compared to a simple simulation using only the best or average parameter set. Additionally, both observed runoff and SWE were used to calibrate the model using a combination of four objective functions which were set to achieve the best possible fit of simulated data with observations (results not shown in the paper). Based on this testing, this multivariable approach improved the model performance especially in snow-dominated catchments and improved both SWE simulations and overall water balance. However, we did not evaluate the mentioned model improvement in detail since this issue is beyond the scope of the paper. The sequential or integrated multivariable model calibration procedure has been used in many other studies and it led to more accurate results in many cases compared to single-variable approach (Bergström et al., 2002; Etter et al., 2017; Rientjes et al., 2013; Seibert, 2000).

The HBV model simulated unrealistic SWE in some snow-rich years in elevation zones higher than 3,000 m a.s.l. (catchments C1, C4, C5, and C6). Here the SWE continually increased during the whole 30 year simulation period. These "snow towers" were probably caused by an overestimation in precipitation in these highest elevation zones and by the degree-day approach used for the snowmelt calculation, which does not reflect radiation input to the entire snowpack energy balance and, probably even more important, snow redistribution is not considered in the modeling (Freudiger et al., 2017). Therefore, these unrealistic amounts of snow melted slower and persisted over the whole warm season. Consequently, snow storage increased to unrealistic values. Since the area above 3,000 m a.s.l represents only 0.2% of the total study area, we assume its influence on water balance and runoff to be negligible. Additionally, the effect of "snow towers" was simulated only for some catchments during the reference period, while future period simulations remained unaffected. The simulated SWE from this highest elevation zone was not included in the analysis presented in Figure 3.

4.2. Climate Change Scenarios and Their Limitation

The climate change scenarios included in data set CH2011 (CH2011, 2011), which has been used in this study, was based on climate model simulations from ENSEMBLE project using A1B, A2, and RCP3PD emission scenarios. A new data set with downscaled projection for Switzerland from the CORDEX project and representative concentration pathways (RCP) is now being prepared (http://www.ch2018.ch/). However, this new data set was not available during the processing of this study.

The CH2011 data set uses a simple delta-change approach to create climate time series by applying predicted corrections. This approach assumes that the interannual variability does not change between the reference period and the three future periods. However, the interannual variability of model outputs might change using the future periods compared to the reference period, since the catchments might have different reactions to changing inputs (especially changes in snowfall fraction and thus SWE and runoff).

Our modeling experiments did not reflect future changes in PET due to changes in air temperature. This was motivated by the temperature-based equations for PET that was used as input to the HBV model. It is known that such types of equations usually overestimate the effect of PET increase, which means that the PET is more sensitive to changes in temperature than is expected in reality (Kingston et al., 2009; Milly & Dunne, 2011; Oudin et al., 2005; Shaw & Riha, 2011). Additionally, explicitly excluding changes in PET into the simulations allowed to better separate the effect of changes in air temperature and snowfall fraction on summer low flows from the effect of a changed PET. Hence, our study results should not be interpreted as a prediction of future low flow changes. Our results instead highlight the effect of air temperature increase and precipitation changes on changes in snow signatures and consequent summer low flow neglecting potential feedbacks from associated changes in PET. Additionally, our analyses were based on the mean prediction in air temperature and precipitation resulting from individual RCMs, but we did not consider the variance in these RCMs. The variance of individual RCMs in estimating future air temperature changes in Switzerland according to the A1B emission scenario, for instance, reached from 1°C to 2.2°C depending on season and future period (CH2011, 2011). The consideration of such variance might be important especially for catchments, where the winter temperatures fluctuate near the freezing point. In such catchments, even slightly different predictions for air temperature might have a large effect on the snowfall fraction and, thus, snow storages.

4.3. Expected Changes in Meteorological and Streamflow Signatures Due to Climate Changes

The results showed a strong decrease in signatures describing snow storage for the three future periods compared to the reference period in all study catchments. This was generally expected and it was also proved by many other studies which focused on this issue in more detail (Beniston et al., 2017; Marty et al., 2017; Sospedra-Alfonso et al., 2015).

It is very difficult to separate the effects of changes in single water balance components on seasonal runoff. Besides the effect of increasing air temperature and thus decreasing snowfall fraction and SWE, the runoff is also influenced by changes in snowmelt timing (both snowmelt onset and melt-out), changes in precipitation (both during the cold season and the warm season), and most importantly, changes in evapotranspiration. We did not explicitly consider changes in PET for future periods in this study as described earlier. Nevertheless, the simulated AET changed due to changes in available moisture. The available moisture changed due to changes in precipitation, but additionally, the HBV model does not simulate any AET from the soil when there is any simulated snow cover.

Based on our results, it seems that both precipitation and AET changes cannot fully explain the changes in monthly or seasonal minimum runoff. Therefore, we conclude that a major part of simulated runoff decrease in the warm season may be explained by changes in snow storages and timing of snowmelt onset and melt-out. This conclusion may also be supported by the fact that the decrease in summer precipitation in high elevation catchments was not as large as in lower elevation catchments (Figure 4). Despite this lower decrease at high elevations, the low flows decreased more at high elevations than at lower elevations (Figure 7). This indicated that at these higher elevations, the low flows were more sensitive to changes in snowfall fraction, SWE, and the shift of snowmelt season toward earlier spring.

The climate change scenarios predict lower precipitation from June to August in catchments above 2,000 m a.s.l. than precipitation in catchments below this elevation. This can be explained by the location of the catchments. Five of the six highest elevation catchments used in this study were located mainly in the eastern part of Switzerland, where annual precipitation is generally lower than in the central and western parts. Therefore, the regional differences between individual catchments are more important in this case than the fact that precipitation increases with elevation. Nevertheless, the climate scenarios showed the decrease in precipitation compared to the reference period in all catchments.

Although catchments with significant glacier runoff regime were not included in the analysis, we cannot completely exclude a minor impact of the glacier meltwater on runoff in case of C6 and C9. However, we assume this influence negligible and we did not find any inconsistencies in streamflow simulations that could be assigned to the glacier influence on catchment runoff.

4.4. Relation Between Snow Storage and Summer Low Flow

The results proved that the influence of snow conditions on summer low flow progressively decreased throughout all future periods due to an increase in air temperature during winter and thus a decrease in snowfall fraction, a decrease in SWE, and the shift of the snowmelt season toward earlier spring at all elevations. The snowfall fraction has an important effect not only on annual discharge (Berghuijs et al., 2014; Speich et al., 2015) but also on summer low flows as documented by Jenicek et al. (2016) in Swiss catchments, Laghari et al. (2012) in Austria, and Godsey et al. (2014) in the western United States. Based on our results we may conclude that summer low flows are significantly sensitive to any SWE changes. This sensitivity might increase problems with water availability in affected regions.

Snow influence on summer runoff was strongly linked to elevation. This "memory effect" from winter to summer was generally longer for catchments above 2,000 m a.s.l. compared to catchments below this elevation and it became progressively shorter for individual future periods. The longer memory effect in these higher elevation catchments was not only related to higher snowpack accumulations but also to a later and longer persisting snowmelt in spring compared to catchments at lower elevations (Jenicek et al., 2016).

The decrease in minimum runoff for the three future periods compared to the reference period is on average much larger than it could be explained by the decrease in precipitation in the warm period and by changes in AET. Additionally, the significant decrease in seasonal Q_{min} from June to August for all future periods compared to the reference period is larger at higher elevations. This decrease in Q_{min} may be mostly related to the decrease in SWE_{max}. The major role of snowmelt was confirmed by the results of the two modeling experiments with changing T_T and S_{FCF}, which simulated the sensitivity of summer low flows to changes in snowpack assuming no changes in total precipitation.

The results presented in Figure 7 indicated that there are years where the decrease in SWE in future periods compared to the reference period is not as large as in other years. Additionally, higher absolute values of low flows occurred which indicate that these higher low flows might be influenced by higher snowpack. However, this occurs only for catchments above 1,500 m a.s.l. The effect of higher summer low flows in snow-rich years was also described by Jenicek et al. (2016) who showed this behavior in the same catchment selection as in this study, using observations since 1971. However, the results presented in our study do not explain the process causality in detail. For example, low flow strongly depends on catchment water storage (Berghuijs et al., 2016; Staudinger et al., 2015) which was not considered in this study. This means that we quantified the relations based on the data we used, but process-based understanding at the catchment scale is limited and has to be further investigated.

4.5. Separation of Changes in Maximum SWE and Snowmelt Season Timing

The results clearly showed that air temperature increase and related decrease in snowfall fraction had an effect not only on the total snow storages (SWE_{max}) but also on the timing of the snowmelt season. This is not surprising and it was described by many other studies (Beniston et al., 2017; Marty et al., 2017; Sospedra-Alfonso et al., 2015). Our results showed that the shift of melt-out toward earlier spring is larger than the shift of snowmelt onset and thus the snowmelt season will be shorter in the future. This was particularly pronounced for elevations higher than 1,500 m a.s.l, where the snowmelt season became shorter by 15–20 days for the last future period compared to the reference period. Additionally, the melt-out shifted by more than 1 month for elevations higher than 1,700 m a.s.l. (30–45 days). Together with the relative decrease in SWE_{max} by up to 75% at elevations less than 2,000 m a.s.l. and by 50% at elevations around 2,500 m a.s.l., it resulted in a strong decrease in late spring and summer low flows.

Modeling experiments with a changing threshold temperature for snow/rain and snowmelt onset (T_T) and with changing amount of snowfall (S_{FCF}) showed higher sensitivity of low flows to changes in these two parameters in catchments above 1,800 m a.s.l. than in catchments below this threshold. The elasticity calculated from modeling experiments with a changing T_T also showed a slightly higher sensitivity than for modeling experiments with a changing S_{FCF}. This was expected since the T_T parameter differentiates not only between snow and rain but it also represents the snowmelt initiation. Therefore, increasing T_T means increasing SWE_{max} (due to higher snowfall fraction and thus higher snowfall) which starts to melt later in the spring (due to later exceeding of air temperature when snow melts). Furthermore, the snowmelt is slower due to a smaller difference between air temperature and threshold snowmelt temperature, which is important for the snowmelt routine. Opposite to T_T, increasing S_{FCF} results only in increasing snowfall and thus increasing SWE_{max} without changes in snowmelt onset and snowmelt rates (the melt-out changes, however, due to higher snow amount which has to be melted). Therefore, a higher sensitivity of low flows to changing T_T was expected. Although this is true for most of the catchments, the differences are rather small. This might indicate that the effect of a decrease in total snow storage on low flows is higher than the effect of an earlier snowmelt. However, more research is needed to make a conclusion.

The HBV model uses the single threshold temperature concept to calculate T_T. Although this threshold value was calibrated separately for each catchment and parameter set, there is still some uncertainty in resulting value which could vary for different meteorological conditions (Harpold et al., 2017; Wayand et al., 2017). It is also important to note, that not only simulated SWE and runoff were affected by changes in T_T and S_{FCF} in our modeling experiments but also simulated AET changed somewhat, because in the HBV model it is assumed that there is no evapotranspiration from the soil or vegetation when there is a snow cover. Therefore, with increasing T_T and S_{FCF}, the simulated AET decreased due to the later melt-out.

5. Conclusions

We found a strong decrease in snow signatures for the three future periods compared to the reference period in all study catchments and elevations. The largest absolute decrease in maximum annual SWE was predicted for elevations from 2,000 to 2,700 m a.s.l. In relative terms, the largest decrease was simulated for elevations below 2,200 m a.s.l. (60–75% for the future period 2070–2099). Above 2,200 m a.s.l., the relative decrease in maximum SWE for the 2070–2099 period is 20–60%. The decrease in maximum SWE was mostly caused by the decrease in snowfall fraction (by 40% at elevations around 2,000 m a.s.l.) due to the increase in air temperature and thus changes in precipitation phase. This increase in air temperature resulted in both earlier snowmelt onset and melt-out. The snowmelt season shifted to be up to 4 weeks earlier. Additionally, the simulated snowmelt season was shortened by 5–20 days depending on the elevation. These large changes in seasonal snow storages will greatly influence the water distribution both in time and space, especially in mountain regions with snowmelt-dominated runoff.

Our results showed that monthly minimum runoff from May to July or August correlated significantly to maximum SWE in catchments with a mean elevation higher than 2,000 m a.s.l. for the reference period. Additionally, this monthly minimum runoff was sensitive to the decrease in maximum SWE. The elasticity reached 0.4–0.9, i.e., a relative decrease in annual maximum SWE by 10% caused a relative decrease in monthly minimum runoff by 4–9%. However, these elasticities strongly decreased for the future periods indicating that snow will be less important in the future to influence low flow. This is especially valid for July and August and for elevations around 1,500–2,000 m a.s.l. For elevations below 1,500 m a.s.l, snow storage did not significantly influence summer low flow in the reference period and thus, large changes in summer low flow due to changes in snow storages are not expected in the future. The results indicated that the decrease in SWE will affect summer low flows more than the earlier snowmelt.

A significant decrease in seasonal minimum runoff from June to August was simulated for all future periods compared to the reference period. This decrease may be mostly related to the decrease in maximum SWE and to the shift of snowmelt season toward earlier spring. Changes in AET and precipitation both played a minor role and cannot explain the changes in both monthly and seasonal minimum runoff. This was also confirmed by two modeling experiments, which simulated a relatively high sensitivity of summer low flows to changes in snowpack assuming no changes in total precipitation. The fraction of runoff originating from snowmelt to total runoff from June to August decreased by more than 50% in the last future period at the highest elevations and almost disappeared at the lowest elevations.

As stated before, it needs to be emphasized that our simulations are based on delta-change approach, which does assume that there is no change in interannual variability and that there might be other processes influencing the catchment response to a temperature increase, which were not captured by our model. However, if the simulated trends are correct, the projected changes in snow storage, snowmelt timing as well as spring and summer low flow would have an extensive impact on spatial and temporal snow and water distribution in mountain regions. The effect of reduced snow storage on low flow might be more critical for regions with seasonal precipitation patterns such as the southwestern United States where winter precipitation (falling as snow in mountains) represent the major part of total annual precipitation. However, our study showed that even in regions with a more uniform precipitation pattern, climate change impacts can be substantial. A large effect on winter tourism and snow making is expected especially at elevations below 2,000 m a.sl. The decrease in snowmelt water volume would affect reservoir management and could cause decreased water availability during the warm period for uses such as hydropower, irrigation and recreation. The decrease in low flow in spring and summer caused by the decrease in snow storages would also affect ecology of river systems. Additionally, summer low flows might become less predictable in snow-dominated catchments in the future because snow, as relatively easy predictable initial condition, will become less important and, thus, low flow predictions will rely more on the less predictable precipitation.

Acknowledgments
Precipitation (RhiresD data product) and temperature (TabsD data product) observational data for this study were made available by MeteoSwiss (for data, contact MeteoSwiss customer service at http://www.meteoswiss.admin.ch/home/about-us/contact.html). Streamflow data were made available by the Federal Office for the Environment (FOEN; for data, contact FOEN costumer service at https://www.hydrodaten.admin.ch/en/publishing-information.html). Input SWE data were prepared by Massimiliano Zappa (massimiliano.zappa@wsl.ch). The climate scenarios data CH2011 may be accessed through the link http://ch2011.ch (CH2011, 2011). Results of all modeling experiments used in this study are publicly available (Jenicek et al., 2017). Support from the Czech Science Foundation (project 13-32133S "Headwaters retention potential with respect to hydrological extremes") and European Union (project EU COST Action 1306, LD15130 "Impact of landscape disturbance on the stream and basin connectivity") is gratefully acknowledged. We also thank Kirsti Hakala for valuable comments regarding climate change data used in this study. Many thanks to Tracy Ewen for improving the language. Finally, we thank Anne van Loon, two anonymous reviewers, and the associated editor for many valuable comments during the review process.

References

Addor, N., Rössler, O., Köplin, N., Huss, M., Weingartner, R., & Seibert, J. (2014). Robust changes and sources of uncertainty in the projected hydrological regimes of Swiss catchments. *Water Resources Research*, *50*, 7541–7562. https://doi.org/10.1002/2014WR015549

Barnett, T. P., Adam, J. C., & Lettenmaier, D. P. (2005). Potential impacts of a warming climate on water availability in snow-dominated regions. *Nature*, *438*(7066), 303–309. https://doi.org/10.1038/nature04141

Barnhart, T. B., Molotch, N. P., Livneh, B., Harpold, A. A., Knowles, J. F., & Schneider, D. (2016). Snowmelt rate dictates streamflow. *Geophysical Research Letters*, *43*, 8006–8016. https://doi.org/10.1002/2016GL069690

Beaulieu, M., Schreier, H., & Jost, G. (2012). A shifting hydrological regime: A field investigation of snowmelt runoff processes and their connection to summer base flow, Sunshine Coast, British Columbia. *Hydrological Processes*, *26*(17), 2672–2682. https://doi.org/10.1002/hyp.9404

Beniston, M., Farinotti, D., Stoffel, M., Andreassen, L. M., Coppola, E., Eckert, N., et al. (2017). The European mountain cryosphere: A review of past, current and future issues. *The Cryosphere Discussions*, 1–60. https://doi.org/10.5194/tc-2016-290.

Berghuijs, W. R., Hartmann, A., & Woods, R. A. (2016). Streamflow sensitivity to water storage changes across Europe. *Geophysical Research Letters*, *43*, 1980–1987. https://doi.org/10.1002/2016GL067927

Berghuijs, W. R., Woods, R. A., & Hrachowitz, M. (2014). A precipitation shift from snow towards rain leads to a decrease in streamflow. *Nature Climate Change*, *4*, 583–586. https://doi.org/10.1038/NCLIMATE2246

Bergström, S., Lindström, G., & Pettersson, A. (2002). Multi-variable parameter estimation to increase confidence in hydrological modelling. *Hydrological Processes*, *16*(2), 413–421. https://doi.org/10.1002/hyp.332

Blahušiaková, A., & Matoušková, M. (2015). Rainfall and runoff regime trends in mountain catchments (Case study area: The upper Hron River basin, Slovakia). *Journal of Hydrology and Hydromechanics*, *63*(3), 183–192. https://doi.org/10.1515/johh-2015-0030

Brahney, J., Menounos, B., Wei, X., & Curtis, P. J. (2017). Determining annual cryosphere storage contributions to streamflow using historical hydrometric records. *Hydrological Processes*, *31*(8), 1590–1601. https://doi.org/10.1002/hyp.11128

CH2011 (2011). *Swiss Climate Change Scenarios 2011*. Zurich, Switzerland: C2SM, MeteoSwiss, ETH, NCCR Climate, and OcCC.

Etter, S., Addor, N., Huss, M., & Finger, D. (2017). Climate change impacts on future snow, ice and rain runoff in a Swiss mountain catchment using multi-dataset calibration. *Journal of Hydrology: Regional Studies*, *13*, 222–239. https://doi.org/10.1016/j.ejrh.2017.08.005

Foster, L. M., Bearup, L. A., Molotch, N. P., Brooks, P. D., & Maxwell, R. M. (2016). Energy budget increases reduce mean streamflow more than snow-rain transitions: Using integrated modeling to isolate climate change impacts on Rocky Mountain hydrology. *Environmental Research Letters*, *11*(4), 44015. https://doi.org/10.1088/1748-9326/11/4/044015

Frei, C. (2014). Interpolation of temperature in a mountainous region using nonlinear profiles and non-Euclidean distances. *International Journal of Climatology*, *34*(5), 1585–1605. https://doi.org/10.1002/joc.3786

Frei, C., & Schär, C. (1998). A precipitation climatology of the Alps from high-resolution rain-gauge observations. *International Journal of Climatology*, *18*(8), 873–900. https://doi.org/10.1002/(SICI)1097-0088(19980630)18:8<873::AID-JOC255>3.0.CO;2-9

Freudiger, D., Kohn, I., Seibert, J., Stahl, K., & Weiler, M. (2017). Snow redistribution for the hydrological modeling of alpine catchments. *WIREs Water*, *4*, e1232. https://doi.org/10.1002/wat2.1232

Godsey, S. E., Kirchner, J. W., & Tague, C. L. (2014). Effects of changes in winter snowpacks on summer low flows: Case studies in the Sierra Nevada, California, USA. *Hydrological Processes*, *28*(19), 5048–5064. https://doi.org/10.1002/hyp.9943

Hanel, M., Vizina, A., Máca, P., & Pavlásek, J. (2012). A multi-model assessment of climate change impact on hydrological regime in the Czech Republic. *Journal of Hydrology and Hydromechanics*, *60*(3), 152–161. https://doi.org/10.2478/v10098-012-0013-4

Harpold, A. A., Kaplan, M. L., Klos, P. Z., Link, T., McNamara, J. P., Rajagopal, S., et al. (2017). Rain or snow: Hydrologic processes, observations, prediction, and research needs. *Hydrology and Earth System Sciences*, *21*(1), 1–22. https://doi.org/10.5194/hess-21-1-2017

Jenicek, M., Seibert, J., & Staudinger, M. (2017). Supplementary data: Modelling of future changes in seasonal snowpack and impacts on summer low flows in Alpine catchments [data set]. *Zenodo.* https://doi.org/10.5281/zenodo.1039954.

Jenicek, M., Seibert, J., Zappa, M., Staudinger, M., & Jonas, T. (2016). Importance of maximum snow accumulation for summer low flows in humid catchments. *Hydrology and Earth System Sciences*, *20*(2), 859–874. https://doi.org/10.5194/hess-20-859-2016

Jonas, T., Marty, C., & Magnusson, J. (2009). Estimating the snow water equivalent from snow depth measurements in the Swiss Alps. *Journal of Hydrology*, *378*, 161–167. https://doi.org/10.1016/j.jhydrol.2009.09.021

Jörg-Hess, S., Fundel, F., Jonas, T., & Zappa, M. (2014). Homogenisation of a gridded snow water equivalent climatology for Alpine terrain: Methodology and applications. *The Cryosphere*, *8*(2), 471–485. https://doi.org/10.5194/tc-8-471-2014

Kingston, D. G., Todd, M. C., Taylor, R. G., Thompson, J. R., & Arnell, N. W. (2009). Uncertainty in the estimation of potential evapotranspiration under climate change. *Geophysical Research Letters*, *36*, L20403. https://doi.org/10.1029/2009GL040267

Klein, G., Vitasse, Y., Rixen, C., Marty, C., & Rebetez, M. (2016). Shorter snow cover duration since 1970 in the Swiss Alps due to earlier snowmelt more than to later snow onset. *Climatic Change*, *139*(3–4), 637–649. https://doi.org/10.1007/s10584-016-1806-y

Laaha, G., Parajka, J., Viglione, A., Koffler, D., Haslinger, K., Schöner, W., et al. (2016). A three-pillar approach to assessing climate impacts on low flows. *Hydrology and Earth System Sciences*, *20*(9), 3967–3985. https://doi.org/10.5194/hess-20-3967-2016

Laghari, A. N., Vanham, D., & Rauch, W. (2012). To what extent does climate change result in a shift in Alpine hydrology? A case study in the Austrian Alps. *Hydrological Sciences Journal*, *57*(1), 103–117. https://doi.org/10.1080/02626667.2011.637040

Langhammer, J., Su, Y., & Bernsteinová, J. (2015). Runoff response to climate warming and forest disturbance in a mid-mountain basin. *Water*, *7*(7), 3320–3342. https://doi.org/10.3390/w7073320

Li, D., Wrzesien, M. L., Durand, M., Adam, J., & Lettenmaier, D. P. (2017). How much runoff originates as snow in the western United States, and how will that change in the future? *Geophysical Research Letters*, *44*, 6163–6172. https://doi.org/10.1002/2017GL073551.

Lindström, G., Johansson, B., Persson, M., Gardelin, M., & Bergström, S. (1997). Development and test of the distributed HBV-96 hydrological model. *Journal of Hydrology*, *201*(1–4), 272–288. https://doi.org/10.1016/S0022-1694(97)00041-3

Luce, C. H., Lopez-Burgos, V., & Holden, Z. (2014). Sensitivity of snowpack storage to precipitation and temperature using spatial and temporal analog models. *Water Resources Research*, *50*, 9447–9462. https://doi.org/10.1002/2013WR014844

Marty, C., Schlögl, S., Bavay, M., & Lehning, M. (2017). How much can we save? Impact of different emission scenarios on future snow cover in the Alps. *The Cryosphere*, *11*(1), 517–529. https://doi.org/10.5194/tc-11-517-2017

Milly, P. C. D., & Dunne, K. A. (2011). On the hydrologic adjustment of climate-model projections: The potential pitfall of potential evapotranspiration. *Earth Interactions*, *15*(1), 1–14. https://doi.org/10.1175/2010EI363.1

Morán-Tejeda, E., López-Moreno, J. I., & Beniston, M. (2013). The changing roles of temperature and precipitation on snowpack variability in Switzerland as a function of altitude. *Geophysical Research Letters*, *40*, 2131–2136. https://doi.org/10.1002/grl.50463

Musselman, K. N., Clark, M. P., Liu, C., Ikeda, K., & Rasmussen, R. (2017). Slower snowmelt in a warmer world. *Nature Climate Change*, *7*(3), 214–219. https://doi.org/10.1038/nclimate3225

Nash, J. E., & Sutcliffe, J. V. (1970). River flow forecasting through conceptual models. Part I—A discussion of principles. *Journal of Hydrology*, *10*(3), 282–290. https://doi.org/10.1016/0022-1694(70)90255-6

Oudin, L., Hervieu, F., Michel, C., Perrin, C., Andréassian, V., Anctil, F., et al. (2005). Which potential evapotranspiration input for a lumped rainfall-runoff model? *Journal of Hydrology*, *303*(1–4), 290–306. https://doi.org/10.1016/j.jhydrol.2004.08.026

Rientjes, T. H. M., Muthuwatta, L. P., Bos, M. G., Booij, M. J., & Bhatti, H. A. (2013). Multi-variable calibration of a semi-distributed hydrological model using streamflow data and satellite-based evapotranspiration. *Journal of Hydrology*, *505*, 276–290. https://doi.org/10.1016/j.jhydrol.2013.10.006

Seibert, J. (2000). Multi-criteria calibration of a conceptual runoff model using a genetic algorithm. *Hydrology and Earth System Sciences*, *4*(2), 215–224. https://doi.org/10.5194/hess-4-215-2000

Seibert, J., & Vis, M. J. P. (2012). Teaching hydrological modeling with a user-friendly catchment-runoff-model software package. *Hydrology and Earth System Sciences*, *16*, 3315–3325. https://doi.org/10.5194/hessd-9-5905-2012

Shaw, S. B., & Riha, S. J. (2011). Assessing temperature-based PET equations under a changing climate in temperate, deciduous forests. *Hydrological Processes*, *25*(9), 1466–1478. https://doi.org/10.1002/hyp.7913

Sospedra-Alfonso, R., Melton, J. R., & Merryfield, W. J. (2015). Effects of temperature and precipitation on snowpack variability in the Central Rocky Mountains as a function of elevation. *Geophysical Research Letters*, *42*, 4429–4438. https://doi.org/10.1002/2015GL063898

Speich, M. J. R., Bernhard, L., Teuling, A. J., & Zappa, M. (2015). Application of bivariate mapping for hydrological classification and analysis of temporal change and scale effects in Switzerland. *Journal of Hydrology*, *523*, 804–821. https://doi.org/10.1016/j.jhydrol.2015.01.086

Staudinger, M., Weiler, M., & Seibert, J. (2015). Quantifying sensitivity to droughts—An experimental modeling approach. *Hydrology and Earth System Sciences*, *19*(3), 1371–1384. https://doi.org/10.5194/hess-19-1371-2015

Staudinger, M., & Seibert, J. (2014). Predictability of low flow—An assessment with simulation experiments. *Journal of Hydrology*, *519*, 1383–1393. https://doi.org/10.1016/j.jhydrol.2014.08.061

Tague, C., & Grant, G. E. (2004). A geological framework for interpreting the low-flow regimes of Cascade streams, Willamette River Basin, Oregon. *Water Resources Research*, *40*, W04303. https://doi.org/10.1029/2003WR002629

Teutschbein, C., Grabs, T., Karlsen, R. H., Laudon, H., & Bishop, K. (2015). Hydrological response to changing climate conditions: Spatial streamflow variability in the boreal region. *Water Resources Research*, *51*, 9425–9446. https://doi.org/10.1002/2015WR017337

Van Loon, A. F., Ploum, S. W., Parajka, J., Fleig, A. K., Garnier, E., Laaha, G., et al. (2015). Hydrological drought types in cold climates: Quantitative analysis of causing factors and qualitative survey of impacts. *Hydrology and Earth System Sciences*, *19*(4), 1993–2016. https://doi.org/10.5194/hess-19-1993-2015

Wayand, N. E., Clark, M. P., & Lundquist, J. D. (2017). Diagnosing snow accumulation errors in a rain-snow transitional environment with snow board observations. *Hydrological Processes*, *31*(2), 349–363. https://doi.org/10.1002/hyp.11002

Wilcox, R. R. (2001). *Fundamentals of modern statistical methods: Substantially improving power and accuracy*. New York, NY: Springer.

Zhang, D., Cong, Z., Ni, G., Yang, D., & Hu, S. (2015). Effects of snow ratio on annual runoff within the Budyko framework. *Hydrology and Earth System Sciences*, *19*(4), 1977–1992. https://doi.org/10.5194/hess-19-1977-2015