



Ensemble projections of future streamflow droughts in Europe

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This discussion paper is/has been under review for the journal Hydrology and Earth System Sciences (HESS). Please refer to the corresponding final paper in HESS if available.

Ensemble projections of future streamflow droughts in Europe

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Received: 8 August 2013 – Accepted: 8 August 2013 – Published: 16 August 2013

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Published by Copernicus Publications on behalf of the European Geosciences Union.

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Abstract

There is growing concern in Europe about the possible rise in the severity and frequency of extreme drought events as a manifestation of global change. In order to plan suitable adaptation strategies it is important for decision makers to know how drought conditions will develop at regional scales. This paper therefore addresses the issue of future developments in streamflow drought characteristics across Europe. Through off-line coupling of a hydrological model with an ensemble of bias-corrected climate simulations (IPCC SRES A1B) and a water use scenario (Economy First), long term (1961–2100) ensemble streamflow simulations are generated that account for changes in climate, and the uncertainty therein, and in water consumption. Using extreme value analysis we derive minimum flow and deficit indices and evaluate how the magnitude and severity of low flow conditions may evolve throughout the 21st century. This analysis shows that streamflow droughts will become more severe and persistent in many parts of Europe due to climate change, except for northern and northeastern parts of Europe. Especially southern regions will face strong reductions in low flows. Future water use will aggravate the situation by 10–30 % in Southern Europe, whereas in some sub-regions in Western, Central and Eastern Europe a positive climate signal may be reversed due to intensive water use. The multi-model ensemble projections of more frequent and severe streamflow droughts in the south and decreasing drought hazard in the north are highly significant, while the projected changes are more dissonant in a transition zone in between.

1 Introduction

Drought is an inherent feature of the water cycle that can occur in virtually all climatic zones. It originates from a temporary aberration of the normal precipitation regime over a large area, but other climatic factors, such as high temperatures and winds, or low relative humidity, can significantly aggravate the severity of the event. Non-climatic

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drivers, such as intensive water use and poor water management, can further exacerbate low-flow conditions in watersheds.

Climate warming is expected to considerably alter the water balance throughout Europe, with higher temperatures resulting in higher potential evapotranspiration as well as in changes in the spatial and temporal distribution of precipitation, including more frequent and persistent dry spells (e.g., Rowell, 2005; Beniston et al., 2007; Christensen and Christensen, 2007; van der Linden and Mitchell, 2009; Nikulin et al., 2011). Hence, with a warmer climate, droughts could become more frequent, severe, and longer-lasting in Europe.

There is a general smaller public awareness of droughts because they usually develop slower and more imperceptibly than other natural hazards, such as floods, tornadoes and earthquakes. Nevertheless, the potential increase in drought hazard with human-induced climate warming has recently become a great concern for the EU (EC, 2007, 2012) given the stresses being placed on water resources and the considerable economical, societal and environmental impacts. In the last two decades, the annual average economic consequences of droughts in Europe amounted to € 6.2 billion (EEA, 2010). The severe drought that hit Southern and Central Europe in the summer of 2003 – with an economic damage of more than € 8.7 billion (EEA, 2010) – showed what the impacts might be if climate change leads to an increase in the frequency and intensity of droughts across Europe (Schär et al., 2004).

There is medium confidence that since the 1950s Southern Europe has experienced a trend to more intense and longer droughts (IPCC, 2012). Stahl et al. (2010) show a trend towards decreasing low flows in most regions of Europe where the lowest mean monthly flow occurs in summer. Some regional studies have confirmed the trend towards reduced low flow conditions in Southern and Eastern Norway (Wilson et al., 2010), the Pyrenees in France (Renard et al., 2008), and the Czech Republic (Fiala et al., 2010). Often, however, detecting a climate change signal in the occurrence and severity of droughts is complicated, as it may be hidden beneath the strong inter-annual to decadal natural climate variability. Moreover, catchments in Europe are often heavily

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disturbed by human influences, which potentially mask the effects of global warming on watershed dynamics. For example, in many river basins in Europe the installment of reservoirs in the course of the 20th century has led to less severe streamflow drought conditions (Svensson et al., 2005). Increasing low flows have also been observed in half of the undisturbed catchments in Finland (Korhonen and Kuusisto, 2010).

Different types of drought can be distinguished, namely meteorological (dryness or precipitation deficit), agricultural (insufficient soil moisture for plant growth), and hydrological drought (deficiency in the bulk water supply). In literature the focus has been largely on the first two aspects. As a result, assessments of future droughts have mainly focused on changes in temperature and precipitation (e.g., Beniston et al., 2007; Blenkinsop and Fowler, 2007; Calanca, 2007; Vidal and Wade, 2009; Sienz et al., 2012; Vidal et al., 2012), or have evaluated changes in soil moisture derived from land surface schemes of climate models (e.g., Burke and Brown, 2008; Sheffield and Wood, 2008; Dai, 2011; Heinrich and Gobiet, 2012).

Sectors such as energy production, river navigation, irrigated agriculture, and public water supply, are directly affected by low surface water levels and limited groundwater storage. To complement studies focusing on meteorological and agricultural drought, this study focuses on changes in hydrological droughts, through the evaluation of anomalies in the low flow spectrum across Europe in view of climate change and water demand projections.

In recent literature, a large number of studies have evaluated the potential impacts of global warming on different components of the hydrological cycle. Relatively few works, however, have focused on changes in low flows (Diaz-Nieto and Wilby, 2005; de Wit et al., 2007; Hurkmans et al., 2010; Majone et al., 2012). Large-scale analyses include the global assessment by Hirabayashi et al. (2008), the works of Lehner et al. (2006) and Feyen and Dankers (2009) for Europe, and of Weiss et al. (2007) for the Mediterranean region. Whereas Hirabayashi et al. (2008) analyzed directly simulated discharges from a GCM, Lehner et al. (2006) and Weiss et al. (2007) applied the monthly-averaged climate change signal of GCMs to observation-based data sets (i.e.,

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a delta change approach), which were then used to drive the WaterGAP global hydrology and water use model (Alcamo et al., 2003; Döll et al., 2003). The coarse temporal and spatial resolution of the climate signal used in these studies does, however, not well reflect potential changes in extreme events at regional and local scales.

5 The pan-European assessment of Feyen and Dankers (2009), on the other hand, employed high-resolution regional climate data from a single regional climate model (RCM) to force a European-wide hydrological model to assess climate-related alterations in the low flow spectrum. They derived low flow characteristics from the simulated streamflow series using extreme value analysis and assessed changes in the magnitude and intensity of streamflow droughts. Results indicated that under the IPCC SRES A2 scenario (see Nakicenovic and Swart, 2000) streamflow droughts will become more severe and persistent in most parts of Europe by the end of this century, except in the most northern and north-eastern regions.

10 These conclusions, however, depend not only on the choice of greenhouse gas emission scenario, but also on the climate model used, as climate model configuration remains the main source of uncertainty in climate projections, especially for European precipitation (Déqué et al., 2007, 2012). Over mid-latitudes in Europe, models even show a disagreement in the direction of change in annual precipitation (see Fig. 11.5 in Christensen et al., 2007). Whereas the magnitude of change in (extreme) precipitation that is simulated by RCMs is, for a large part, determined by the driving global model, the regional model formulation influences the spatial pattern (Fowler et al., 2007). This is because the large-scale circulation patterns within RCMs largely depend on the lateral boundary conditions from their driving GCM, influencing not only the mean precipitation changes but also the extremes. Considering uncertainties in low precipitation extremes, Blenkinsop and Fowler (2007) demonstrated considerable dependency on the driving GCM for future projections, particularly for drought frequency. Following from this, an ensemble-based framework considering multiple driving climate scenarios will provide a more robust estimation of the future changes in streamflow drought hazard.

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ALADIN-RM5.1, HIRHAM5, CLM, RACMO2, HadRM3, REMO) (see Table 1). Accordingly, an ensemble of 12 climate experiments with daily temporal resolution and cell size of ca. 25 km, covering the period 1961–2100 and forced by the SRES-A1B scenario (Nakicenovic and Swart, 2000), was used.

The climate experiments were corrected for bias in the precipitation and minimum, average, and maximum temperature fields using the Quantile Mapping (QM) method (Piani et al., 2010a, b; Dosio et al., 2012). Several techniques to correct potential bias in precipitation and temperature have been recently developed in literature based on different transfer functions and theoretical assumptions (Teutschbein and Seibert, 2010; Themeßl et al., 2011). We implemented the QM method because it showed higher performances to correct bias in high-resolution regional climate model data compared to other methods (Themeßl et al., 2011). Dosio and Paruolo (2011) showed that the QM bias correction procedure drastically improved the agreement between simulated and observed climatology from the E-OBS dataset (Haylock et al., 2008). Moreover, it has been shown to yield improvements in both the upper and lower tail of the probability distribution functions (PDFs) of temperature and precipitation, with significantly positive effects on the accuracy of simulated extreme climatic and hydrological events (Dosio and Paruolo, 2011; Rojas et al., 2011).

2.2 Water use scenarios

To evaluate the potential effects of water consumption on streamflow drought indices a spatially distributed quantitative water use scenario from WaterGAP3 (Aus der Beek et al., 2010; Flörke et al., 2012, 2013; Schaldach et al., 2012) was used. WaterGAP3 consists of different sub-models to determine water withdrawal and consumption in different sectors (domestic, tourism, energy, manufacturing, irrigation and livestock). The water use scenario was taken from the SCENES project (Kämäri et al., 2008; Kok et al., 2011) which aimed at developing and analyzing a set of comprehensive water-related scenarios for Europe through a participatory process. Four comprehensive scenarios were developed: Economy First (EcF), Fortress Europe (FoE), Policy Rules (PoR), and

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Sustainability Eventually (SuE). The scenarios include consistent projections of the main drivers such as total population, GDP, thermal electricity production, agricultural production as well as information on technological changes. The EcF scenario was selected as it is the most coherent with the IPCC SRES A1B. It is characterized by a globalised and liberalized economy pushing the use of all available energy sources accompanied by a marked agricultural intensification. The adoption of new technologies and water-saving consciousness is low resulting in an increasing water demand of all water-related sectors. Only water ecosystems providing ecological goods and services for economies are preserved and improved.

Within SCENES water uses for the different sectors are modeled on an annual basis, except for the irrigation water use which has a monthly temporal resolution. Note that the SCENES scenarios run until 2050. For the remaining period it was assumed that water consumption remains unchanged from 2050 onwards.

2.3 Hydrological scenarios

River discharge simulations for different climate experiments (see Table 1) were obtained using the LISFLOOD model (van der Knijff et al., 2010). Being a fully-distributed and physically-based hydrological model developed for large-scale impact assessment studies, LISFLOOD simulates the spatial and temporal patterns of catchment responses as a function of spatial information on meteorology, topography, soils, and land cover. It is a GIS-based hydrological model where processes such as infiltration, water consumption by plants, snowmelt, freezing of soils, surface runoff and ground-water storage are explicitly accounted for at the grid level. Spatial properties for soils, vegetation types, land uses, and river channels constitute the basic input information to set up a LISFLOOD run, whereas data on precipitation, air temperature, potential evapotranspiration, and evaporation from water bodies and bare soil surfaces, are the main meteorological drivers. Potential evapotranspiration and evaporation rates are calculated from vapor pressure, wind speed, radiation (solar + thermal), albedo, and average, minimum, maximum and dewpoint temperature through the off-line LISVAP

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pre-processor based on the Penman–Monteith equation (van der Knijff, 2008). We recognize the potential importance of reservoirs and flow regulation for hydrological low-flow analysis, particularly in smaller catchments. However, such structures have not been implemented in our assessment due to the lack of suitable information on dams, artificial reservoirs and their current and future operation. As such, the actual magnitudes of the low flow measures derived herein more reflect conditions in undisturbed catchments. It can be argued that, unless considerable alterations in flow regulation take place to mitigate the severity of extreme low flow conditions, the relative changes in low flow measures presented should be less affected by not accounting for regulation. For a more detailed description of the processes and equations of LISFLOOD, as well as of its calibration, we refer the reader to van der Knijff et al. (2010) and Feyen et al. (2007, 2008).

Prior to forcing the LISFLOOD model, the climate simulations were re-gridded to the 5 km LISFLOOD grid employing a nearest neighbor approach on the basis of the centre points of the 25 km grid cells of the RCMs. The water use data were similarly re-gridded from their 5 arc minute grid to the LISFLOOD grid. The WaterGAP3 monthly (for irrigation) and annual (for the other sectors) water consumption data were equally distributed over the days in each month or year, respectively, and accounted for in LISFLOOD as a daily loss term. LISFLOOD was then run with a daily time step for a simulation period between 1961 and 2100. As such, for each experiment (climate ensemble member combined with or without water use) 140 yr of daily discharges were produced at each river pixel. To analyze changes over this period, time slices of 30 yr were considered, further herein referred to as control period (1961–1990), 2000s (1981–2010), 2020s (2011–2040), 2050s (2041–2070), and 2080s (2071–2100). For each period a flow duration curve (FDC) was derived in each river pixel from the 30 yr simulated discharge time series. FDC represents the percentage of time that river flow is likely to exceed some specified value and has been used herein as the basis for the calculation of the low-flow indices detailed below.

2.4 Indices of streamflow drought

Various drought indices have been developed to monitor and quantify droughts. For an extensive overview on low-flow and drought indices and their derivation we refer the reader to Smakhtin (2001), Tallaksen and van Lanen (2004) or Mishra and Singh (2010). In this work we follow the approach of Feyen and Dankers (2009) and focus on two important aspects of a drought, namely its magnitude and its persistence through time.

Firstly, we analyze low flows through the magnitude of the river discharge, expressed here by the 7 day minimum flow (q_{min}) at several recurrence intervals. We apply a 7 day averaging to cancel the day-to-day fluctuations in river flow. From the smoothed discharge series the annual minima are selected, through which a Generalized Extreme Value (GEV) distribution is fitted using the Maximum Likelihood method (Gilleland and Katz, 2005). From the fitted GEV distribution q_{min} values for different return periods ranging between 2 and 100 yr are derived in each river pixel.

Secondly, we consider the development in time of drought events by evaluating deficit characteristics of periods in which discharge stays below a threshold flow. Deficits can be evaluated in terms of their run duration (length of event) and severity (cumulative deficit or negative run sum) (Smakhtin, 2001). However, given the often strong correlation between drought durations and deficit volumes (Woo and Tarhule, 1994) and considering that the latter are a more effective measure of the magnitude of water shortage relevant for operational water management (Tallaksen and van Lanen, 2004), we focus only on deficit volumes (def). The threshold for evaluating deficits can be defined as a percentage of the mean flow or as an exceedance frequency of the FDC. We opt for the latter, such that everywhere in Europe discharge time series fall below the threshold an equal number of days, but allowing deficit volumes to vary according to location-specific conditions. As a balance between representing low flow conditions and assuring sufficient events for extreme value fitting (Tallaksen et al., 1997; England et al., 2004; Fleig et al., 2006) we apply the 80 % exceedance frequency of the

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FDC (further referred to as Q_{80}) as a threshold. We note that for the future time slices, deficits are evaluated against the threshold from the control period (1961–1990). Using the ML method, in each river cell a Generalized Pareto (GP) distribution was then fitted through the partial duration series of deficit volumes representing the shortfalls below the threshold. From the fitted distribution deficit volumes for recurrence intervals ranging between 2 and 100 yr were derived.

The selection of drought events with the threshold method using a daily time resolution is, however, plagued by two problems. Firstly, when in periods of prolonged low flow conditions the flow shortly exceeds the threshold level, the large event is split in two or more smaller mutually dependent events. Secondly, shortfalls below the threshold of very short duration yield very small deficit volumes that may cause instability in the extreme value distribution fitting (Engeland et al., 2004; Tallaksen and van Lanen, 2004; Fleig et al., 2006). Tallaksen et al. (1997) evaluated three different pooling procedures to reduce the occurrence of minor events and reduce mutual dependency. On the basis of their results, and similar as for the low-flow indices described above, we apply a moving average (MA) procedure with a 7 day averaging window to the discharge time series prior to selecting the events, which removes minor droughts and pools mutually dependent droughts. As not all minor events were excluded after the 7 day MA procedure, in each river cell we additionally removed all events with a deficit smaller than 0.5 % of the maximum deficit in that cell, following Zelenhasic and Salvai (1987).

2.5 Streamflow regimes in Europe

Streamflow regimes across Europe vary strongly due to the large variability in climatologic conditions and local factors that influence the hydrological response, such as the presence of aquifers, variations in soil properties and land cover. In regions with a cold climate, typically winter and summer droughts can be differentiated (e.g., Fleig et al., 2006; IPCC, 2012). In winter, most water is trapped as snow and ice, often resulting in the lowest flows seen throughout the year. Hence, by applying an annual analysis there

is the risk that low flows in the frost-free season, originating from negative imbalances in precipitation, are not accounted for, or that the sample of events used for extreme value fitting find their origin in different physical processes. The latter also violates the underlying theory of frequency analysis, namely that the events are considered to be drawn from an independent and identically distributed (iid) random variable.

Following Hisdal et al. (2001) and Feyen and Dankers (2009) we distinguish between a nonfrost and frost season, where the latter is defined for each river pixel as the period of the year in which the monthly average temperature in the upstream area drops below zero in at least 23 out of 30 yr (reference length of time slices). By averaging over the upstream area we avoid labeling river pixels as nonfrost when in large parts of the upstream catchment water is still stored as ice or snow. This is especially the case for downstream river reaches in areas with pronounced topography, such as for the major streams that drain from the Alps (e.g., Rhine, Rhone, or Po rivers).

For intermittent and ephemeral streams the data series of annual minima may contain several zero values, hence discontinuous probability distribution functions need to be applied for inferring low flow probabilities. Analysis of the LISFLOOD simulations showed, however, that for catchments with an upstream area larger than 1000 km² the time series of annual minima did not contain any zero flow values, and thus can be considered to be perennial under current and future climatic conditions. We therefore limit the analysis to river basins with an upstream area exceeding 1000 km², hereby excluding ephemeral rivers. Even if the latter (headwater and lower order streams) may be particularly sensitive to climate change, as analyzed in a comprehensive review by Brooks (2009), to be properly quantified they would require a finer hydrogeomorphological characterization than the one provided in the large-scale approach presented herein. Furthermore, we argue that river basins with an upstream area exceeding 1000 km² are representative enough to explore the impact of climate changes on future hydrological droughts at continental scale.

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2.6 Uncertainty in low-flow projections

2.6.1 Climate uncertainty

Climate models are the most robust tools available to generate consistent climate change projections. However, they are still source of considerable uncertainties due to the incomplete, missing or incorrect representation of some physical processes and approximated parameterizations (e.g., Katz, 2002; Murphy et al., 2004; Déqué et al., 2012). One of the crucial issues emerging from recent studies (see, e.g., Christensen and Christensen, 2007; van der Linden and Mitchell, 2009) is that different climate experiments may still show large variations in the simulated variables, especially for precipitation. This may be translated to the impact models, often resulting in considerable climate-induced variability in impact estimates.

Model uncertainties can be partly resolved using an ensemble-based framework where simulations from different climate models are used to drive the impact assessment model. Projections based on a multi-model ensemble can be considered more reliable than projections produced by single models alone, as the multi-model average or median can be expected to outperform individual ensemble members, thus providing an improved “best estimate” projection (IPCC, 2007). On the other hand, it should be noted that a multi-model ensemble, especially when the sample is relatively small, may still be affected by extreme or outlier ensemble members. Also, over the ensemble, errors in one process or parameterization may be compensated by errors in other processes or parameterizations (e.g., Murphy et al., 2007). The weighting of climate simulations as an approach to combining climate information is subject to considerable debate (e.g., Christensen et al., 2010; Coppola et al., 2010; Déqué and Somot, 2010). Any weighting method depends on subjective choices about the metrics and combining procedure into an overall weight for the individual models. Such factors will determine the spread in the climate and impact estimates and add an additional layer of uncertainty. Therefore, climate simulations of the different ensemble members have been equally weighted when summarizing the low-flow projections.

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In order to address uncertainty related to climate change projections we first quantified the consistency in the projections of streamflow drought indices. Here “consistent” is interpreted as the agreement amongst the 12 ensemble members in terms of showing a decrease or increase in low-flow measure of at least 5% with respect to the control period.

The significance of the changes in low flow indices in the ensemble was evaluated by comparing the difference between the mean of the 12 ensemble members of the control period and the mean in the scenario period against the variability within the ensemble for the respective periods. As the variance in the future periods is not necessarily the same as in the control run, we used an adaptation of Student’s t test, which is commonly known as Welch’s t test (Welch, 1947; see also Von Storch and Zwiers, 1999, p. 113). Here, the test statistic can be approximated with a normal t distribution, where the appropriate degrees of freedom are estimated from the data. If the resulting p value is smaller than a predefined significance level, e.g. 5% ($\alpha = 0.05$), the ensemble mean in the future time slice is said to be significantly different from that in the control period given the climate uncertainty.

2.6.2 Extreme value fitting uncertainty

The aim of extreme value analysis is to find a parametric model for the tail of the data generating process, then to fit this model to the extreme observations and use it for extrapolation beyond the observed data. Extreme value modeling typically faces the problem of data scarcity, or that the fitting is based on a relatively small number of observations. This introduces uncertainty in the estimation of the return levels that depends on the quantity of data in relation to the degree of extrapolation. Here this uncertainty has been appraised by applying the profile-likelihood method (Coles, 2001) on the return levels of the GEV and GP distributions and estimating the corresponding 95% confidence intervals. The profile-likelihood has been demonstrated to be more robust and able to better capture the usual asymmetric nature of the confidence inter-

vals than other conventional methods (e.g., delta method) (Coles, 2001; Beirlant et al., 2004).

We recognize that errors in the conceptualization, parameterization, and in the calibration of the hydrological model further contribute to the overall uncertainty in the modeling chain. Some studies have recognized this layer of uncertainty to be of secondary importance (e.g., Wilby, 2005; Najafi et al., 2011), while others regard it as important (Bastola et al., 2011; Wada et al., 2013). Moreover, the quantification and modeling of environmental, social, and policy drivers of water use, such as population dynamics, land use changes, agricultural, industrial, energy and environmental policies, as well as economic and technological developments, are inherently uncertain (Kok et al., 2011). Therefore actual water consumption could deviate from the SCENES projections used herein.

3 Results and discussion

3.1 Validation of low flow simulations

A large set of 446 gauging stations across Europe (see Fig. 1) – for which long enough daily data time series were available (30 yr of records for the majority of stations) – was employed to validate simulated low flow indices for the control period (1961–1990). A preliminary screening procedure was implemented to avoid the inclusion of stations with ephemeral conditions and multi-year observational gaps. The validation network covers a wide range of hydro-climatic regimes in Europe and varies in the size of contributing upstream areas from ca. 1000 km² to ca. 810 000 km². As shown in Fig. 1, the validation stations are not evenly distributed across Europe, with a high density of stations located in western and central parts of Europe and hardly any in Italy and South-Eastern Europe. At 338 stations the number of years with a frost season in the control climate was at least 23 out of 30; hence they were used for validation in both the nonfrost and frost season analysis.

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Regional climate simulations of the control period (1961–1990), although bias corrected such that they statistically conform to observed climate, do not reproduce historical weather. This does not allow a day-to-day or event-to-event comparison. Instead, we evaluate the accuracy of the LISFLOOD low flow simulations using statistical measures calculated over 1961–1990. More specifically, we employ the coefficient of determination (r^2), model efficiency (EF), and percent bias (PBIAS), defined as

$$r^2 = \left\{ \frac{\sum_{i=1}^N (Y_{\text{obs},i} - \bar{Y}_{\text{obs}}) (Y_{\text{sim},i} - \bar{Y}_{\text{sim}})}{\left[\sum_{i=1}^N (Y_{\text{obs},i} - \bar{Y}_{\text{obs}})^2 \right]^{0.5} \left[\sum_{i=1}^N (Y_{\text{sim},i} - \bar{Y}_{\text{sim}})^2 \right]^{0.5}} \right\}^2 \quad (1)$$

$$\text{EF} = 1 - \left[\frac{\sum_{i=1}^N (Y_{\text{obs},i} - Y_{\text{sim},i})^2}{\sum_{i=1}^N (Y_{\text{obs},i} - \bar{Y}_{\text{obs}})^2} \right] \quad (2)$$

$$\text{PBIAS} = 100 \cdot \left[\frac{\sum_{i=1}^N (Y_{\text{sim},i} - Y_{\text{obs},i})}{\sum_{i=1}^N Y_{\text{obs},i}} \right] \quad (3)$$

where $Y_{\text{obs},i}$ and $Y_{\text{sim},i}$ are the observed and simulated low flow index at station $i = 1, \dots, N = 446$, respectively, and the horizontal bar denotes averaging over all stations. The coefficient of determination (r^2) describes how much of the observed dispersion is explained by the prediction. It ranges between 0.0 and 1.0 with better agreement for higher values. The model efficiency (EF) determines the relative magnitude of the simulated error variance compared to the observed data variance. It ranges between minus infinity and 1.0, with higher values for increased model performance. The percent bias (PBIAS) measures the average tendency of the simulations to be larger or smaller than observations. It ranges from -100 to $+100$ with low-magnitude values (close to 0) indicating accurate model prediction, whereas positive (negative) values indicate overestimation (underestimation). Each of the goodness-of-fit measures has specific

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pros and cons which have to be considered for model evaluation (e.g., Krause et al., 2005; Moriasi et al., 2007). Statistics on the dispersion (r^2) and variance (EF) of the observed vs. simulated data and the model tendency to under/over-estimate (PBIAS) complement each other by providing different reliability measures and serve to assess the overall modeling performance.

Figure 2 presents the performance of the 12 ensemble members to reproduce 7 day average minimum flows in the control period. Results show a good performance across the climate simulations with r^2 and EF values ranging between 0.94–0.96 and 0.81–0.95, respectively. Two common features can be observed for all members of the ensemble. Firstly, at some stations the simulated minimum flows deviate strongly from those derived from observations, a behavior that is more pronounced with decreasing catchment size. This relates to multiple sources of uncertainty that are more dominant in small basins. These include, among others: intrinsic limitations of RCMs to reproduce small-scale processes; conceptual approximations in the LISFLOOD model, its input data and parameterization; not fully accounting for reservoir/lakes and flow regulation in the modeling setup; and measurement errors at river gauging stations. The lower model performance for smaller catchments implies that future projections of changes in low flow conditions presented herein should be interpreted more cautiously compared to those obtained for larger catchments. We also note that by coupling off-line RCM simulations with the hydrological model, as well as by correcting the bias only in temperature and precipitation, the energy and water balance are not necessarily preserved. Notwithstanding the latter, Rojas et al. (2011) showed the strong improvement in LISFLOOD performance after bias correction.

Secondly, there is a general tendency to underestimate observed minimum flows, as expressed by the negative values of PBIAS (ranging between -13 and -40.5%). This is most likely related to the underestimation of the low-end percentiles of the bias corrected precipitation, and a potential overestimation of the number of dry days obtained from the fitting of the transfer functions (Dosio and Paruolo, 2011; Rojas et al., 2011).

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Figure 2 also shows that models with a large bias not necessarily show the lowest performance based on the other statistics; hence the ranking of the ensemble members in terms of performance depends on the metric considered. This suggests that different ensemble members may outperform other members with respect to simulating different aspects of the low flow spectrum, which reinforces the idea of a multi-model framework as the basis for streamflow drought impact assessment due to climate change. This also supports the findings of Lenderink (2010), who explored different metrics of extreme daily precipitation to conclude that there is no metric that guarantees an objective and precise ranking or weighting of climate models.

Figure 3 presents the ensemble-averaged performance for different low flow indices, where Fig. 3a corresponds to the average performance of the individual members presented in Fig. 2. Noteworthy is the strong negative bias for the average deficit volumes (Fig. 3b) compared to average minimum flows (Fig. 3a). This implies that LISFLOOD driven by the climate simulations tends to underestimate flow deficits (or less severe droughts), but also, albeit to a lesser extent, minimum flows (or more severe droughts). What seems a conflicting discrepancy in bias between the simulation- and observation-based statistics (i.e., model simulates less vs more severe droughts depending on index), suggests that the underestimation of the low-end percentiles of the bias corrected precipitation and the overestimation of the number of dry days have a relatively higher impact on the low-end percentile range Q_{80} – Q_{99} than on more extreme low flows.

A comparison between the statistics of the fitted extreme value distribution (here exemplified by minimum flows and deficits with a 20 yr recurrence interval) shows nearly equally good performance for the minima and deficits (Fig. 3c, d). The decreases in PBIAS observed for fitted extreme values with respect to the averages may be explained by the fact that the former includes an optimization process that minimizes discrepancies between observed and simulated values, thus reducing the bias. High values of the statistical measures confirm that LISFLOOD simulations are fairly robust in capturing extreme streamflow droughts occurring in the nonfrost season.

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From Fig. 3e, f it can also be seen that LISFLOOD driven by control climate simulations has problems in reproducing runoff and base flow in the frost season, resulting in a larger tendency to underestimate the low flow indices in comparison to the nonfrost season. This is due to a combination of several factors, including (1) conceptual and parameter errors in the snow and frost modules of LISFLOOD that affect drainage in the cold season; (2) an incorrect parameterization of the groundwater storage due to bias in the observed winter precipitation (see Goodison et al., 1998; Yang et al., 2001) that was used in the LISFLOOD calibration; (3) too low temperatures during intermittent melt events, resulting in more water that remains stored as snow and less base-flow generation; and (4) not properly accounting in the LISFLOOD setup for storage release from reservoirs to guarantee minimum flow requirements for example for hydropower production. This may also induce artificially higher thresholds (Q_{80}) than those derived from the simulated flow duration curve, yielding at a large number of stations larger observed deficit volumes compared to those simulated, especially under severe drought conditions.

From the validation, we can conclude that for the nonfrost season LISFLOOD driven by a multi-model ensemble of bias-corrected regional climate simulations is able to reproduce streamflow droughts, expressed by the average annual minima and deficits (Fig. 3a, b) and the corresponding 20 yr events (Fig. 3c, d), reasonably well across the wide range of climatic and hydromorphological conditions of Europe. For the frost season, LISFLOOD simulations are less reliable due to the reasons described above, hence these results should be interpreted more cautiously.

3.2 Changes in meteorological forcing and water consumptions

To better understand the processes affecting future low flow characteristics across Europe we first summarize the projected changes in the main driving climatic variables. Figure 4 shows for Europe the ensemble-average changes by the 2080s compared to control climate (1961–1990) in temperature and precipitation for the nonfrost and frost season. Because streamflow at a given location depends on the hydroclimatological

conditions over the upstream river basin, these maps show average changes over the contributing area rather than the change at the location itself. Figures 5 and 6 (top 2 rows) present ensemble-average alterations in these variables throughout the year at a selection of stations that span the range of climatic and hydrological conditions in Europe (see Fig. 1).

Figure 4 shows that temperature in the nonfrost season is projected to increase all over Europe, with the strongest warming (up to 4 °C increase by the end of this century) in southern parts of Europe. This is confirmed by the station plots in Figs. 5 and 6, where the increase in summer temperature at southern stations (Fig. 6, stations Beaucaire, Pontelago, Lugo and Seros) is more pronounced than at those located further north (Fig. 5, stations Langnes, Isohaara and Dau Gavpil). In cold regions, the rise in temperature during the frost season is projected to be higher than in the nonfrost season (e.g., stations Langnes, Isohaara and Dau Gavpil). This is conform to other works based on different RCMs (e.g., Christensen et al., 2007), which showed the strongest increase in temperature during winter for northern parts of Europe. Due to warming, the frost season, defined here as the period in the year for which in at least 23 out of 30 yr the monthly average temperature drops below zero, will shorten (e.g., stations Langnes, Isohaara, Dau Gavpil and Neuhausen), or disappear completely (e.g., stations Boguslawi and Harsova). Consequently, the spatial extent of areas with a frost season in the control climate, shaded in light gray in Fig. 4, is projected to shrink (shift to higher latitudes and altitudes) considerably in the course of the 21st century. The absence of a permanent frost period due to global warming will strongly affect the hydrological cycle and ecosystem functioning in these regions.

Average precipitation in the nonfrost season is projected to decline in Southern Europe, with decreases as high as 30 % in the most southern regions, to rise by 10–20 % in Northern Europe, and to remain relatively stable in a transition zone in between. Comparison of the stations plots in Fig. 6 shows that in the most southern parts of Europe (see stations Lugo and Seros) drying is much stronger in spring compared to summer, whereas further north (see stations Montjean and Beaucaire, Ponte

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Lago) equally strong reductions are observed in summer precipitation. Note that the average maps over the nonfrost season may mask some of the inter-annual changes. At Kingston station (Fig. 5), for example, average precipitation is projected to slightly increase, but summer precipitation will decrease. In the frost season average precipitation is projected to strongly increase over most parts of Northern Europe (e.g., stations Langnes, Isohaara and Dau Gavpil, Neuhausen in Fig. 5), which is in line with other studies using different climate models and emission scenarios (see e.g., Christensen and Christensen, 2007; Räisänen and Eklund, 2012).

Figure 7 shows total annual water withdrawals aggregated to river basin for the control period and the corresponding expected changes in the 2050s according to the EcF scenario from SCENES. This figure summarizes current and future water use to better capture possible human-induced intensification of streamflow droughts. Most intense water abstractions can be observed in the UK, the Benelux countries, Germany and Northern Italy with more than 100 mm of annual water consumption. Lower consumption (< 25 mm) is observed on Northeast Europe, Western France and to some extent Hungary as well as Croatia. The Iberian peninsula and countries of Eastern Europe show intermediate total water withdrawals. Water withdrawals are expected to increase in the 2050s by more than 50 % in Western, Eastern and Northern Europe mainly due to the growing thermal electricity production in the energy sector and an increasing manufacturing gross value added (Flörke et al., 2011). At the same time, irrigation water requirements will play a major role in Northern Iberian Peninsula and Northern Italy due to an intensification of crop production in combination with increasing temperatures induced by climate changes and will lead to an increase of ca. 25 % in total water abstractions. Slight decreases in future total annual water withdrawals can be observed in some river basins in Denmark, Southern Iberian Peninsula, Southern Italy and Greece. Additional details on changes in water consumptions for the different sectors can be found in Flörke et al. (2011).

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3.3 Projections of future streamflow droughts

3.3.1 Frost season

The bottom row in Figs. 5 and 6 shows how the changes in meteorological forcing affect the streamflow dynamics across Europe. In stations with a frost season (e.g., stations Langnes, Isohaara and Dau Gavpil, Neuhausen in Fig. 5), low flows in this period are projected to augment considerably, hence winter droughts to become less severe. In warmer and wetter winters, a smaller portion of precipitation will be temporarily stored as snow or ice, resulting in increased flows in the cold season. This is in line with the findings by Räisänen and Eklund (2012) of a decrease in long-term mean snow water equivalent throughout the 21st century in Northern Europe.

The magnitude of change in the low flow spectrum in the frost season is positively correlated with the magnitude of change in precipitation and temperature in the upstream basin. However, it also depends on the hydroclimatologic conditions prior to the start of the frost season, especially for the more rare events as they reflect not only changes in the frost season but rather imbalances over longer time spans. In these regions summer and to a lesser extent also autumn precipitation is projected to decline, resulting in subsurface storages that are relatively smaller at the start of the frost season. The extreme or very rare minimum flows are therefore expected to show a relative increase that is less pronounced than for more moderate low flow conditions. These findings are fully consistent with the results of Feyen and Dankers (2009). Given that droughts during the cold season will pose considerably less serious problems, the further analysis focuses only on the nonfrost season.

3.3.2 Future streamflow minima

For the nonfrost season, changes in 7 day minimum flows with recurrence intervals ranging from 2 up to 100 yr were derived from the GEV distributions fitted through the annual minima series for the control and future time slices. Figure 8 (left column)

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regions are typically a summer or autumn phenomenon. Minimum flows with relatively short recurrence intervals, which reflect the water balance in the preceding months, are strongly impacted by the pronounced decrease in summer and autumn precipitation. The more rare events, on the other hand, reflect imbalances in precipitation over longer time spans, in which the reduced precipitation input over summer and autumn is counterbalanced by increased subsurface storages at the start of the summer season due to elevated precipitation amounts in winter and spring.

In northern parts of Europe an opposite signal is observed in the non-frost season and minimum flows prevalently increase (or become less severe) in time. Scandinavia and the Baltic countries will experience a general increase in 20 yr minimum flows up to 20 % – in some inland tributaries up to 40 % – by the end of the 21st century. This is a result of the increase in precipitation that outweighs the effects of increased evapotranspiration demands with higher temperatures. Here the change to less severe droughts is more pronounced for rarer events due to the considerable increase in precipitation during the frost season, which after melting of snow in spring yields larger volumes of water stored in the subsurface at the start of the summer season. In some high-latitude areas, however, the projected rise in winter precipitation may not necessarily result in thicker snow packs, which in combination with earlier snowmelt may result in relatively stable minimum flows in the frost-free season (e.g., some southern parts of Sweden and along the west coast of Norway).

3.3.3 Future streamflow deficits

Changes in 7 day streamflow deficits with return periods ranging between 2 and 100 yr were obtained from the fitted GP distributions for deficit volumes in both the control and scenario time slices. We present changes in 20 yr deficits (def) (Fig. 8, right column) for river pixels with upstream catchment size exceeding 1000 km². Note that in this case red colors indicate an increase in flow deficits, which implies more severe shortfalls below the threshold (Q_{80} of control period) or more severe droughts.

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In the nonfrost season flow deficits are projected to become more severe in most of Europe, except for northern and north-eastern regions. In many regions of the Mediterranean – including the Iberian Peninsula, Italy and Balkan – as well as parts of Eastern Europe – including Bulgaria and Romania – 20 yr deficit volumes are expected to increase by 50 % by the 2020s. These regions will experience continued drought intensification up to the end of the century, with deficit volumes (below Q_{80} of the control period) increasing by up to 80 % by the 2080s. From the 2050s onwards most of France, the Netherlands, Belgium, UK and the Alpine regions will also be prone to more severe streamflow deficits (increases between 20 % and 50 %). Although not shown here and in agreement to what was observed for the minimum flows, events with shorter recurrence intervals show stronger increases in deficit volumes (mainly in southwestern parts of Europe). It should be noted, however, that the extrapolation error when fitting the extreme value distribution beyond the length of the time series increases with recurrence interval. Hence, projections for higher return periods are more prone to uncertainty, which also explains the somewhat more scattered pattern in the changes of deficit volumes for higher return periods in some regions.

In North-Eastern Europe, including the Baltic countries, flow deficits in the nonfrost season show a general declining trend, with reductions in deficit volumes up to 60 % and more by the end of this century. North-western parts of Europe display a general decreasing trend in deficit volumes, but locally (e.g., along the Norwegian west coast, areas in Southern Sweden) the changes show higher spatiotemporal variability with sparse rivers showing an opposite tendency in the signal of change. This mixed pattern is due to the combined effects of a general increase in precipitation and reduced snowmelt contribution caused by less accumulation of snow in winter. Depending on the relative magnitude of these processes in future time slices, streamflow deficits in this region may either become more or less severe.

A comparison of the projected changes in deficit volumes (Fig. 8, right column) with those in minimum flows (Fig. 8, left column) for the different time slices shows that the relative alterations in flow deficit volumes are more pronounced than those in the

magnitude of minimum flows. Some regions where minimum flows are projected to stay more or less stable or slightly increase may even show increasing trends in deficit volumes (e.g., Northern France, Benelux, UK and Ireland, Hungary and sub-regions of Romania). This suggests that streamflow droughts will develop longer in time, or become more persistent and that the flow spectrum in the range of Q_{90} – Q_{70} is more affected than very extreme low flows. This implies that streamflows in the scenario climate may show more frequent and intense shortfalls below the Q_{80} threshold of the control period, even when extreme minimum remain relatively stable. Again, this is due to the fact that extreme low flows are affected by hydroclimatologic conditions over longer time windows, in which the negative impact of drying in the nonfrost season is partly offset by wetter winters. More frequently occurring low-flow conditions, on the other hand, reflect short term imbalances in precipitation and are thus largely determined by the strong precipitation reduction in summer season. Similarly, rarer deficit volumes (i.e., with higher recurrence intervals) are therefore less affected at mid latitudes.

3.3.4 Impact of water consumption on streamflow droughts

The results described above only show the effect of climate change on low flow characteristics. The impact of increased water consumption on low flow characteristics is presented in Fig. 9, which shows the ensemble-average change between the 2080s and the control period in 20 yr minimum flows and deficit volumes when accounting for both climate change and water consumption (Fig. 9a, b). It is worth noting that the 7 day-average streamflow accounting for both climate change and water use is also included in Figs. 5 and 6 as a red line in the bottom row panels. Note that to account for water consumption, LISFLOOD was coupled on a daily time step with WaterGAP3 for the whole simulation period 1961–2100, hence the changes in low flows with respect to the control period reflect the combined effects of alterations in climate and consumptive water use. The lower panels of Fig. 9 show for each drought index the difference

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between the ensemble-average changes driven by changes in climate and water use and those accounting only for the climatic ones for the 2080s.

Intensive water consumption as projected by the EcF scenario will further aggravate streamflow droughts in many regions of Europe. It will negatively affect both minimum flows and deficit volumes in Central, Western and Eastern Europe due to the projected increases in water abstractions (compare with Fig. 7). In some regions where no or slightly positive changes in low flow conditions are induced by climate change, increasing water consumption will reverse this trend and lead to more severe streamflow droughts. This behavior is most notable in the Benelux countries, Western Germany, Northwestern France (see station Montjean in Fig. 6), and localized parts in the UK (see station Kingston in Fig. 5), and in Central and Eastern European countries (Slovakia, Czech Republic, Hungary and Romania). Water use abstraction will exacerbate low flow conditions by ca. 10–30 % over the Mediterranean regions (see e.g. stations Seros, Lugo, Ponte Lago and Beaucaire in Fig. 6). This suggests that even in front of a relative reduction in water abstractions (compare with Fig. 7), simulations performed over the EcF scenario project more pronounced negative changes in minimum flows than those performed accounting only for the climate change. In regions with positive signal of warming on low flow conditions, such as the Scandinavian peninsula and Baltic countries, intensive water use may reduce future low flows, but not sufficiently to offset the positive signal.

3.4 Uncertainty in projections of streamflow droughts

Figure 10 presents the consistency amongst the 12 ensemble members in projecting a decrease (left panels) or increase (right panels) in minimum flows in future time slices with respect to the control period. Consistent patterns are found for deficits and are not shown here for brevity. For the 21st century, the majority of hydrological simulations depict consistently increasing streamflow drought conditions over the Iberian Peninsula, France, Italy, UK and Southern Balkans. On the other hand, models also agree well about the decrease in streamflow droughts in Northeastern Europe, Scandinavia,

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the Baltic countries, and northern parts of Poland. A more mixed pattern with higher variability in low-flow regimes across streamflow drought simulations is evident mainly over the transition zone across central Europe and the Carpathians. A common feature for the minimum flow and deficit indicators is the improved agreement in time between projections of the ensemble members in the aforementioned areas, thus, suggesting a stronger signal-to-noise ratio as time proceeds.

Figure 11 shows the resulting p value from the Welch's t test for 20 yr minimum flows and deficit volumes. Significance levels appear to be inversely correlated with the ensemble-average absolute magnitude of change in minimum flows and deficit volumes (see Fig. 8). As the magnitudes of drying in the south and wetting in the north increase with time, the changes become also more robust, confirming a higher signal-to-noise ratio. By the end of the 21st century, for nearly all river pixels where the ensemble-average change in low flow measure exceeds 5 %, the changes are statistically significant at the 5 % level ($\alpha = 0.05$). In those areas where the ensemble-average changes in low flow measures are small, higher p values are found, thus suggesting a weaker signal-to-noise ratio. This indicates that the models tend to show less agreement (see also Fig. 10) about the direction (and magnitude) of change in the transition zone between clearly defined regions with increasing (south) and decreasing drought hazard (north). Although not shown here, the same behavior was observed when water consumption is accounted for.

The high consistency amongst the different climate member in projecting streamflow drought changes suggests that the decadal-scale internal climate model variability, which may partially or completely obscure the climate signal in extreme events, is of secondary importance. Low flows largely depend on imbalances in precipitation over longer (i.e., monthly to seasonal) timescales, rather than on single events as is the case for example for floods (Rojas et al., 2012). At these scales the changes in rainfall that determine the drought hazard are well established and more consistent between climate models.

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where all ensemble member low flow simulations clearly show less severe streamflow drought conditions in the future. In the nonfrost season, more overlap of the spread can be observed, but still a clear increase ensemble average low flows is observed at these stations.

Figure 13 shows the additional uncertainty arising from the fitting compared to that related to climate change. Left panels visually exemplifies the uncertainty sources in the EV analysis of annual minimum flows and deficits derived from the hydrological simulations obtained from the different members of the climate ensemble (Fig. 13a, f, respectively). The dark gray area reflects the uncertainty arising from employing alternative climate simulations to force LISFLOOD (CC uncertainty), where the thick black line represents the ensemble average of the fitting distribution (r_l). The light gray area, on the other hand, reflects the additional uncertainty arising from the extreme value fitting (Fit uncertainty) and is expressed by the 95 % confidence intervals on the return levels averaged on the twelve hydrological simulations. Marginal uncertainty of fitting (MUF) is calculated as the ratio of uncertainty fitting over the total uncertainty – including climatic and fitting sources – for a given return period. Right panels of Fig. 13 show the relationship between the magnitude of CC and Fit uncertainty (on the x- and y-axis, respectively) for all stations (minimum flows and deficit are shown in Fig. 13b–e and g–j, respectively). Point colors refer to the magnitude of return levels (r_l), whereas $avg(MUF)$ represents the marginal uncertainty of fitting averaged over all stations.

GEV fitting of annual minimum flows introduces an additional source of uncertainty – minor compared to that of climate change – ranging from 9 to 13 % (Fig. 13b–e). Both Fit and CC uncertainties tend to decrease for increasing return period, as effect of the lower bounded shape of the GEV distribution that tends to converge towards inferior variability in the upper tails. Interestingly, fitting uncertainty tends to decrease faster compared to climate change uncertainty along increasing return period, especially for stations with lower return levels, as exemplified by the corresponding reduction in average MUF. This suggests that the variability of projections of most low flow extreme events is more influenced by the climate model variability than the fitting confidence.

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Projections of minimum flows indicate an overall increase in magnitude of uncertainties for both climate and fitting. However, fitting uncertainty is expected to play a secondary role with respect to the climate uncertainty as shown by the reduction in average MUF in 2080s compared to the control period.

Uncertainty in the GP fitting plays a more important role for the deficit volumes than for the minimum flows, as quantified by higher average MUF ranging from 60 % to 38 % (Fig. 13g–j). The GP fitting to the deficit volumes is susceptible to errors as partial duration series of deficits below a threshold often contain, even after smoothing the discharge series as applied here (see Sect. 2.4.), a large number of minor droughts that distort the inference of the scale and especially the shape parameter of the GP distribution. This may result in fitted GP distributions that rapidly overshoot the underlying data used for fitting. Even if the magnitude of both climate and fitting uncertainties increase in future time slices, the average MUF tend to decrease in future projections of deficits, consistently to what is observed for the minimum flows. This confirms that extreme value fitting of annual minima and deficit volumes may introduce additional sources of uncertainty in projections of streamflow droughts playing, however, a secondary role with respect to the climate uncertainty.

4 Conclusions

Here we have assessed the implications of global warming and water consumption on low flow conditions in Europe. We first generated an ensemble of streamflow scenarios from 1961 to 2100 that account for future climate developments – and the climate model uncertainty therein – under the IPCC SRES A1B scenario, as well as for changes in consumptive water use under a coherent scenario (Economy First, FP6 SCENES project). In a second step, streamflow drought indices were derived using extreme value analysis and changes between different 30 yr windows were analyzed.

Our analysis has led to the following three main conclusions.

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ditional sources of uncertainty in projections of streamflow droughts, however, it seems secondary with respect to the climate uncertainty. Even hydrological uncertainty – here not accounted for – may further increase the variability in the low flow projections as suggested by the considerable discrepancy between large-scale hydrological models in the evaluation of drought propagation (see Van Loon et al., 2012).

The expected changes in streamflow drought presented herein largely agree with those based on the analysis of projections of climate and land surface scheme variables (e.g., Heinrich and Gobiet, 2012; Sienz et al., 2012). The strong signals in increasing drought severity that show high statistical significance indicate that many sub-regions in Europe will increasingly face water stress throughout the 21st century. Southern regions of Europe, which suffer already most from limited water availability, will be affected hardest, but also other regions in Western, Central and Eastern Europe will likely face more stringent drought conditions. The strong reductions in water availability in low flow periods that will happen more frequently will increase the competition for water amongst sectors (e.g., energy, agriculture, households). This may pose unforeseen challenges for policy makers and water managers in the regions affected to map out adequate adaptation strategies in order to minimize the socio-economic and environmental impacts of these changes.

Acknowledgements. We acknowledge past and present colleagues at the JRC who assisted over the years in developing and calibrating LISFLOOD. Data used in this work was funded by the EU-FP6 projects ENSEMBLES (contract no. 505539) and SCENES (contract no. 036822), whose support is gratefully acknowledged. We thank A. Dosio for the bias correction of the ENSEMBLES climate simulations and M. Zambrano-Bigiarini for assistance in R coding.

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Table 1. Climate simulations used to drive LISFLOOD in the period 1961–2100.

Model	Driving GCM	RCM	Institute	Acronyms
1	HadCM3Q16 ^a	RCA3.0	The Community Climate Change Consortium for Ireland	C4I-RCA-HadCM3
2	ARPEGE	ALADIN-RM5.1	Centre National de Recherches Météorologiques, Meteo France	CNRM-ALADIN-ARPEGE
3	ARPEGE	HIRHAM5	Danish Meteorological Institute	DMI-HIRHAM5-ARPEGE
4	BCM	HIRHAM5	Danish Meteorological Institute	DMI-HIRHAM5-BCM
5	ECHAM5-r3 ^b	HIRHAM5	Danish Meteorological Institute	DMI-HIRHAM5-ECHAM5
6	HadCM3Q0 ^a	CLM	Swiss Federal Institute of Technology	ETHZ-CLM-HadCM3
7	ECHAM5-r3 ^b	RACMO2	The Royal Netherlands Meteorological Institute	KNMI-RACMO2-ECHAM5
8	HadCM3Q0 ^a	HadRM3Q0	Hadley Centre for Climate Prediction and Research, UK	METO-HadRM3-HadCM3
9	ECHAM5-r3 ^b	REMO	Max-Planck-Institute for Meteorology, Germany	MPI-REMO-ECHAM5
10	BCM	RCA3.0	Swedish Meteorological and Hydrological Institute	SMHI-RCA-BCM
11	ECHAM5-r3 ^b	RCA3.0	Swedish Meteorological and Hydrological Institute	SMHI-RCA-ECHAM5
12	HadCM3Q3 ^a	RCA3.0	Swedish Meteorological and Hydrological Institute	SMHI-RCA-HadCM3

^a Represent three versions of the HadCM3 model with perturbed parameterization impacting the simulated climate response sensitivities: Q0 (reference), Q3 (low-sensitivity) and Q16 (high-sensitivity) (Collins et al., 2006). ^b Represent one run of the ECHAM5 model using three different sets of initial conditions defined as “-r1”, “-r2”, and “-r3” (Kendon et al., 2010).

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Fig. 1. Location of the 446 gauging stations used to evaluate simulated streamflow drought indices. Filled black circles show validation stations used in Figs. 6, 7 and 12.

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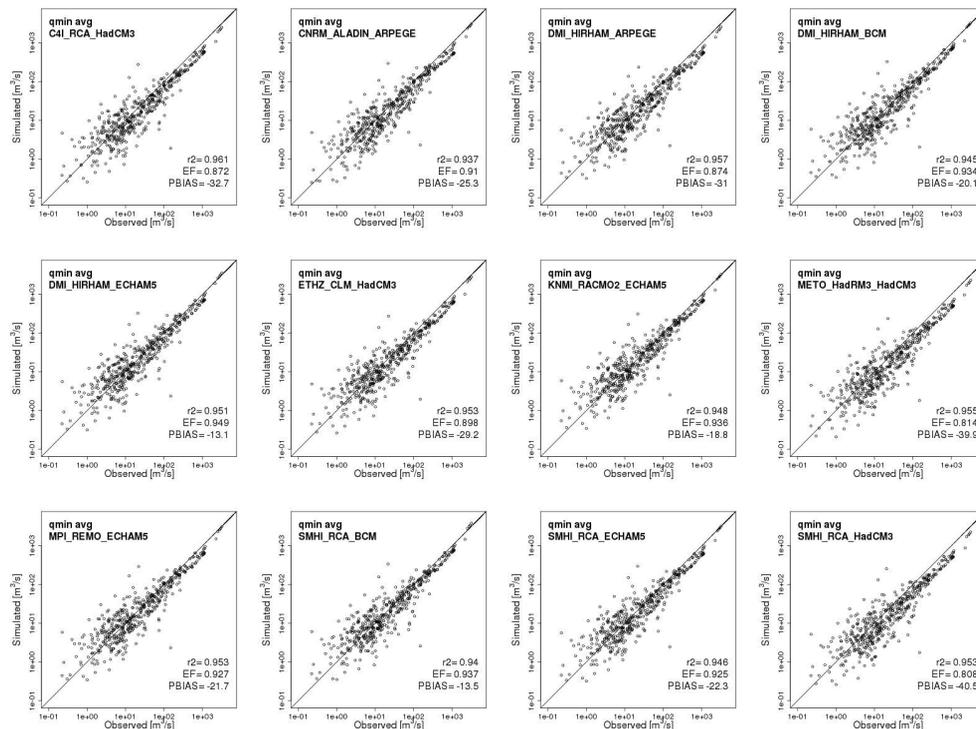


Fig. 2. Observed vs. simulated minimum flows for the control period (1961–1990) at each of the 446 stations depicted in Fig. 1 based on the hydrological simulations driven by the 12 climate experiments listed in Table 1.

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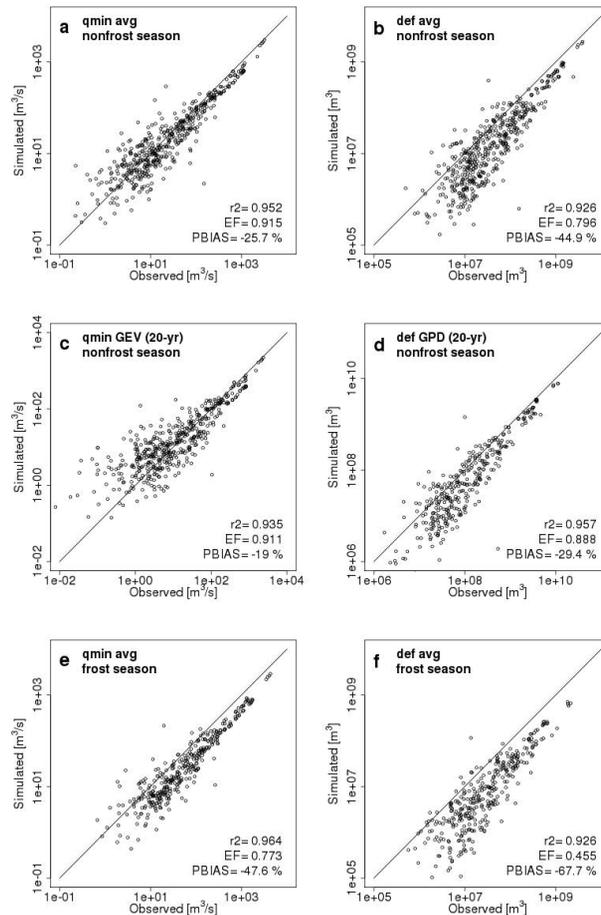


Fig. 3. Observed vs. ensemble-averaged simulated streamflow drought indices for the control period (1961–1990) at each of the 446 stations depicted in Fig. 1.

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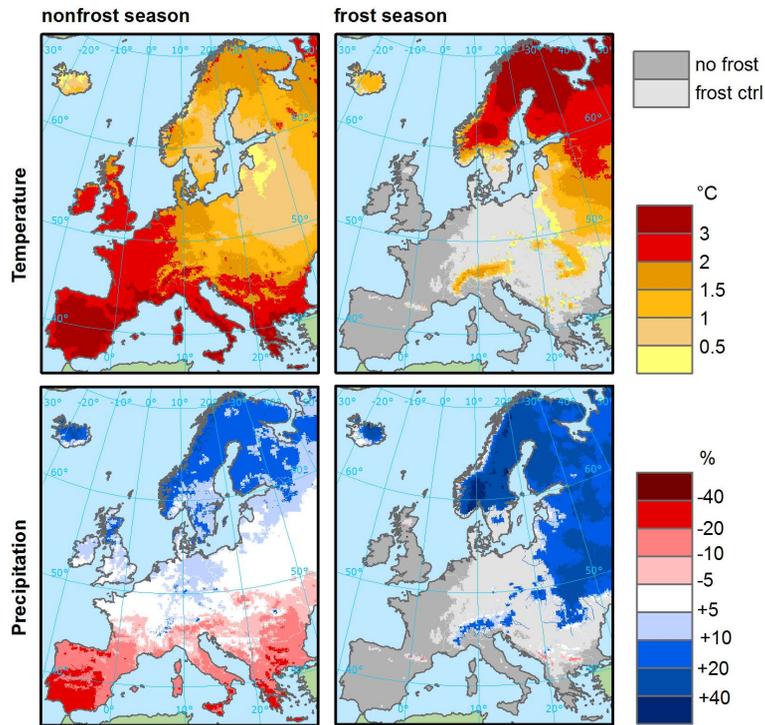



Fig. 4. Ensemble-averaged changes in average temperature (top row) and precipitation (bottom row) between 2080s and control period for nonfrost (left column) and frost (right column) season. The change at each location reflects the average change over the upstream area contributing flow to that location. In the frost season panels areas shaded on light gray represent regions with a frost season in the control climate and without a frost season in the ensemble-average scenario climate, while dark gray areas indicate regions with no frost season both in the control and scenario periods.

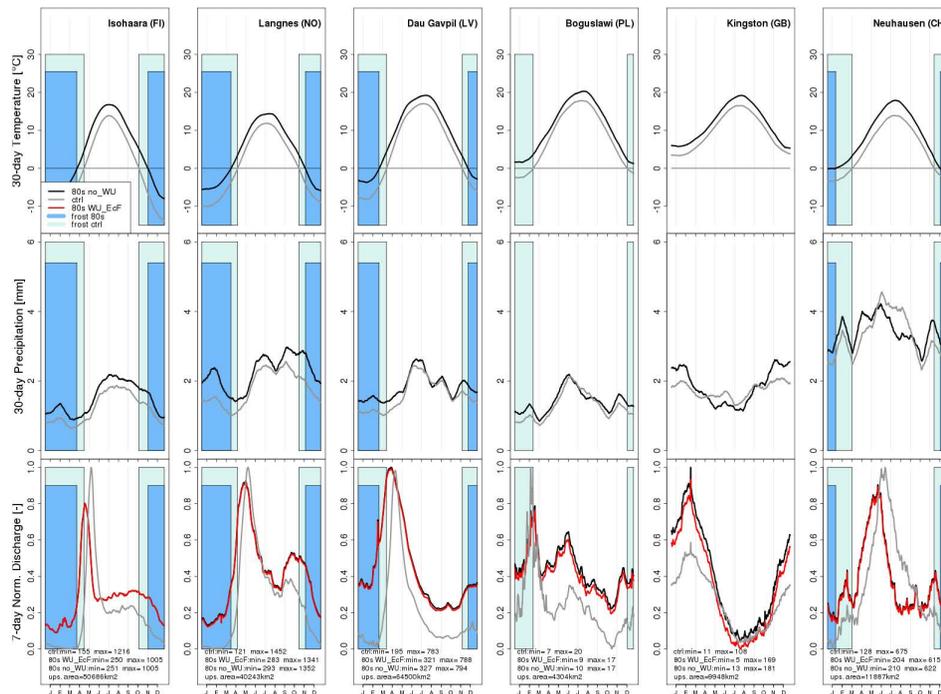


Fig. 5. Inter-annual dynamics in temperature (top row), precipitation (middle row) and 7 day average streamflow (bottom row) at selection of stations (see Fig. 1) in the control period (light gray lines) and 2080s (black line). Temperature and precipitation reflect 30 day upstream averages. The red line for streamflows (bottom row) reflects the scenario accounting for water use. Blue shaded areas indicate the length of the frost season in both periods (control = light blue, 2080s = dark blue).

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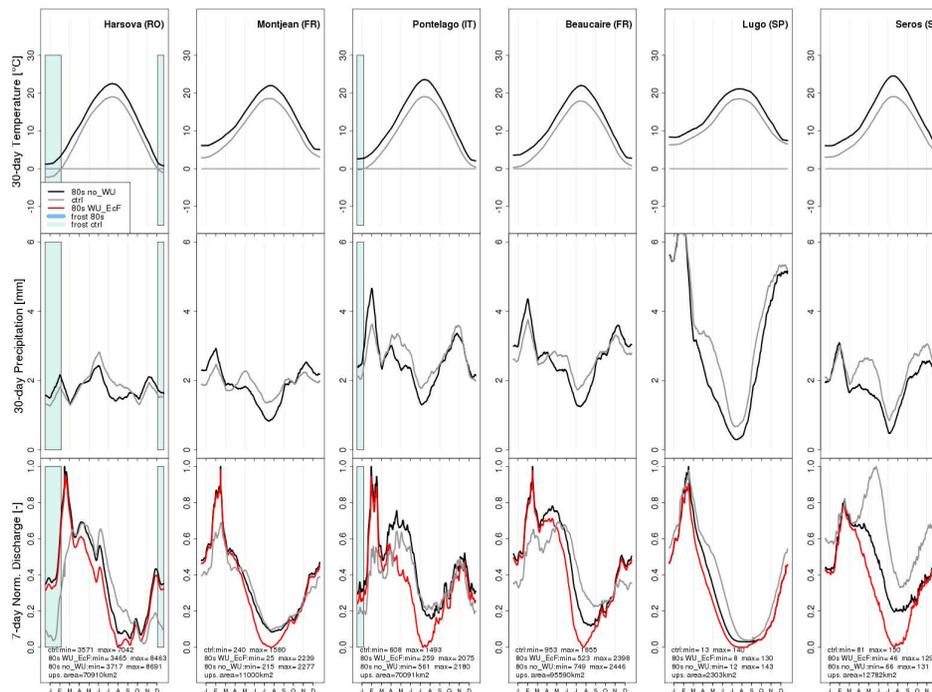


Fig. 6. As Fig. 5, for different stations.

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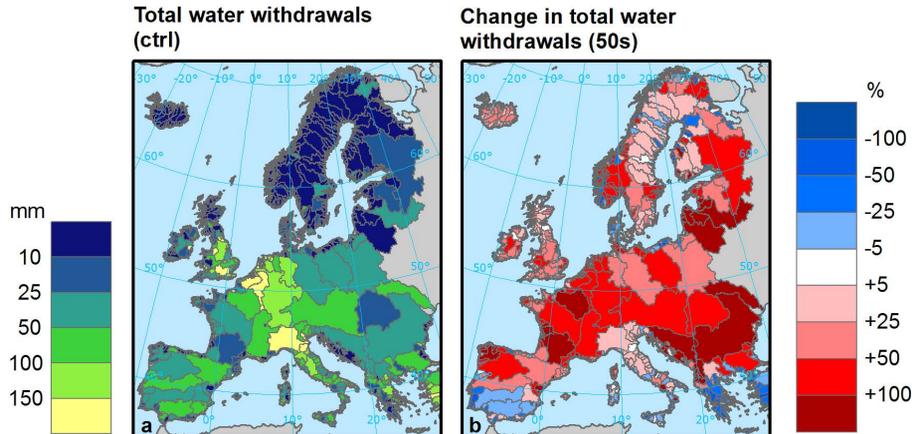


Fig. 7. Total annual water withdrawals aggregated to river basin scale for the control period **(a)** and the corresponding expected changes in the 2050s according to the EcF scenario from SCENES project **(b)**.

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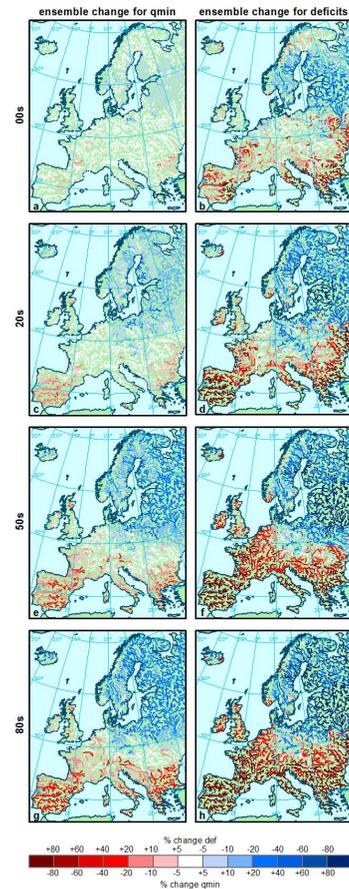


Fig. 8. Ensemble-average change in 20 yr return level minimum flow (left) and deficit volumes (right) due to only climate change between the corresponding time slices and the control period (1961–1990).

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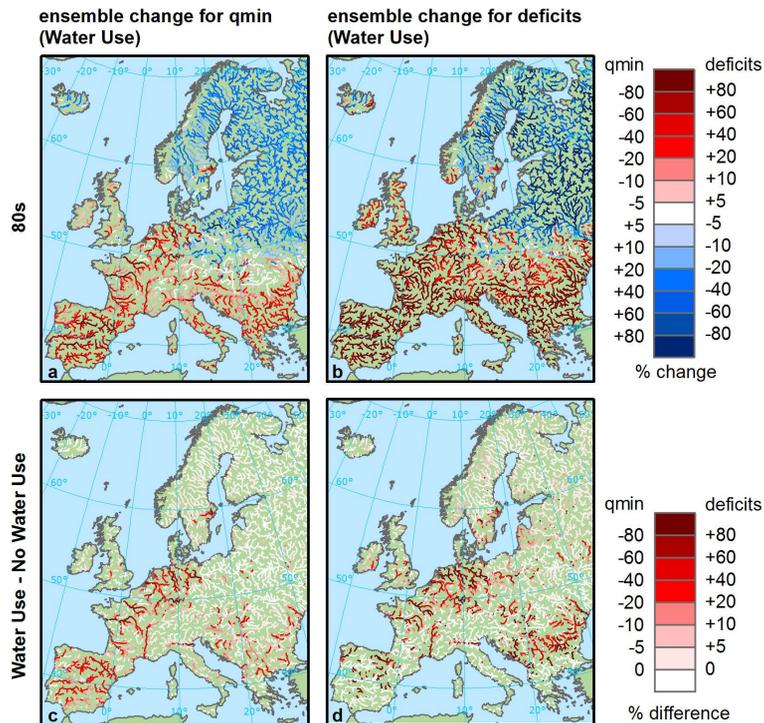


Fig. 9. Ensemble-average change in 20 yr minimum flow and deficit volumes due to climate change and water consumption between the 2080s and the control period (top row) and corresponding differences with ensemble-average changes driven only by climatic drivers (bottom row).

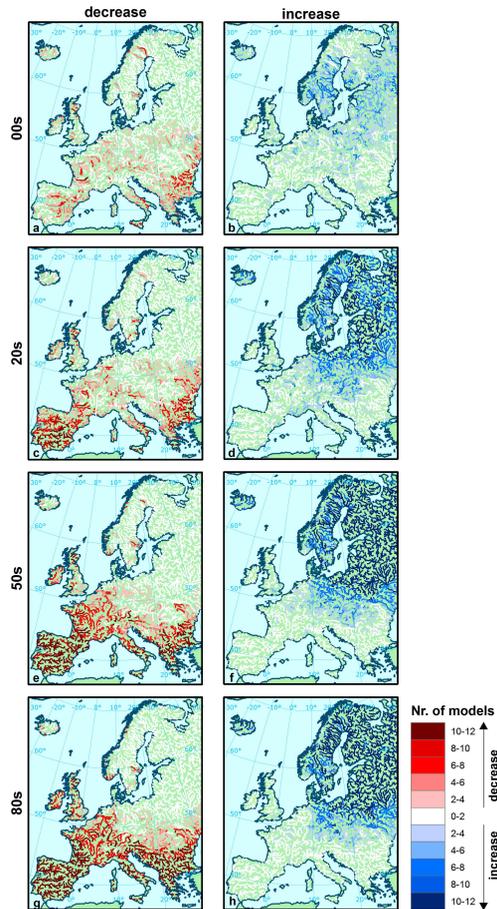


Fig. 10. Consistency of the streamflow drought simulations for different time slices indicated by the number of simulations (out of 12) agreeing in a decrease (left) or increase (right) of more than 5% in minimum flows.

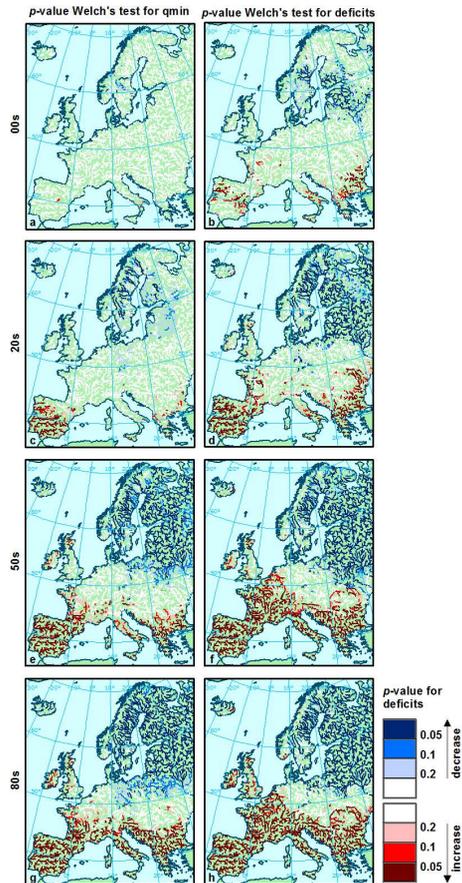


Fig. 11. p value to test significance of the average change in minimum flows (left) and deficit volumes (right) between the corresponding time slices and the control period (1961–1990). p values are obtained on the basis of the Welch's t test.

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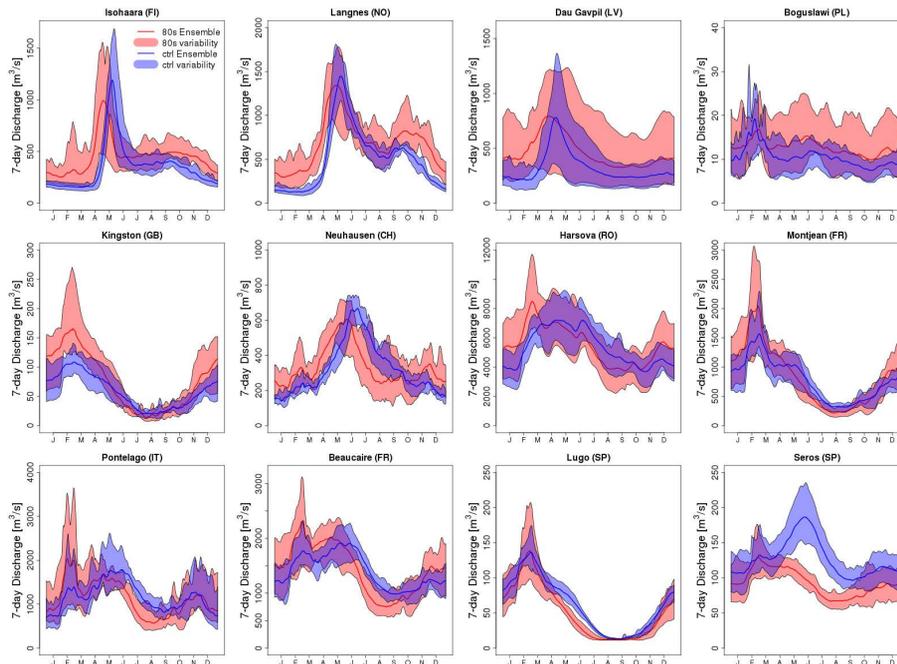


Fig. 12. Inter-annual dynamics in simulated 7 day streamflow for control period and 2080s for selected stations (see Fig. 1). The thin blue and red lines represent the ensemble averages for the control and 2080s, respectively. The corresponding shaded areas show the spread amongst ensemble members within the respective period.

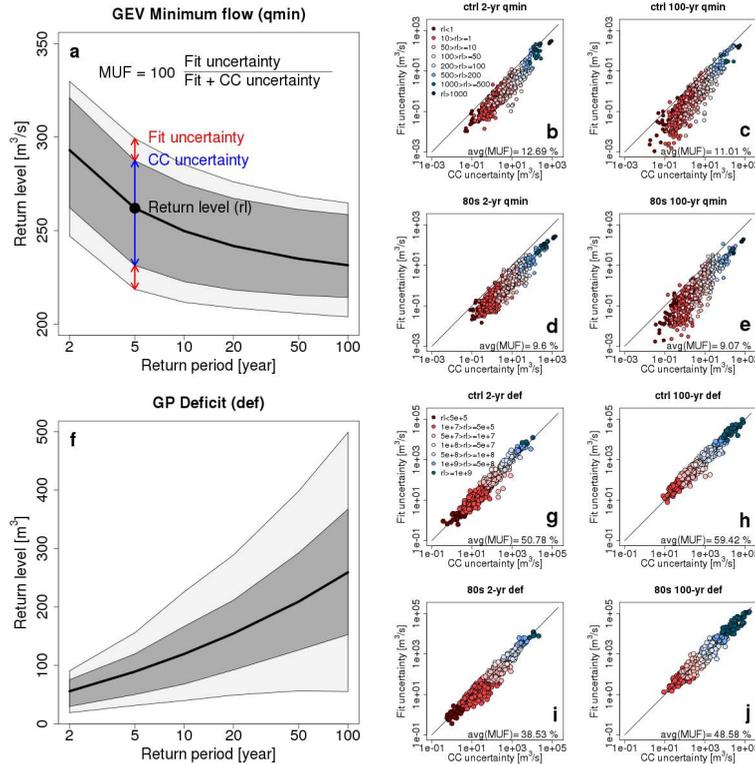


Fig. 13. Relationship between the magnitude of climate change (CC) and EV fitting (Fit) uncertainties over the 446 gauging stations (see Fig. 1) for the control and 2080s periods and for the 2 yr and 100 yr return periods. Colors of points refer to the magnitude of return levels (rl). GEV results for minimum flows shown in (a) through (e), while GP results shown in (f) through (j).

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