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## Projections of future floods and hydrological droughts in Europe under a +2°C global warming

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### ► To cite this version:

Philippe Roudier, Jafet C. M. Andersson, Chantal Donnelly, Luc Feyen, Wouter Greuell, et al.. Projections of future floods and hydrological droughts in Europe under a +2°C global warming. *Climatic Change*, 2016, 135 (2), pp.341-355. 10.1007/s10584-015-1570-4 . hal-01235952

**HAL Id: hal-01235952**

<https://hal.sorbonne-universite.fr/hal-01235952v1>

Submitted on 1 Dec 2015

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22 Germany, France and North of Spain. North of this line, floods are projected to decrease in most of  
23 Finland, NW Russia and North of Sweden, with the exception of southern Sweden and some coastal  
24 areas in Norway where floods may increase. The results concerning extreme droughts are less robust,  
25 especially for drought duration where the spread of the results among the members is quite high in  
26 some areas. Anyway, drought magnitude and duration may increase in Spain, France, Italy, Greece, the  
27 Balkans, south of the UK and Ireland. Despite some remarkable differences among the hydrological  
28 models' structure and calibration, the results are quite similar from one hydrological model to another.  
29 Finally, an analysis of floods and droughts together shows that the impact of a +2°C global warming will  
30 be most extreme for France, Spain, Portugal, Ireland, Greece and Albania. These results are particularly  
31 robust in southern France and northern Spain.

32

### 33 1. Introduction

34 In Europe, freshwater resources are of crucial importance for many sectors including agriculture,  
35 hydropower generation, cooling water for power plants and domestic and industrial water supply. At  
36 the same time, water can have a direct impact on safety and livelihoods through floods that can lead to  
37 disastrous human and economic losses (e.g. the 2013 floods in Central Europa resulted in a loss of more  
38 than €12bn<sup>1</sup>). During the last decades, several studies have underlined that water resources and  
39 especially river flows have had strong variations across Europe (Kovats et al., 2014) due to climate,  
40 water extractions and land use change (Sterling et al., 2013). Even if the relative share of these three  
41 driving factors is difficult to assess over the past, it is clear that the strong climate changes expected in  
42 Europe for the 21<sup>st</sup> century will have a significant impact on river flows (Jiménez Cisneros et al., 2014).

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<sup>1</sup> <http://www.munichre.com/en/media-relations/publications/press-releases/2013/2013-07-09-press-release/index.html>

43 Thus, a first step in order to make relevant mitigation and adaptation policies is to develop a clear  
44 picture of the potential future stream flows extremes.

45 Changes in river flow extremes at a +2°C global warming are currently of central interest as this is  
46 the global target defined by policymakers to lower international greenhouse gases emissions (European  
47 Commission, 2007). Therefore, in this study, we do not select a specific time period; rather we focus on  
48 defining the hydrological impacts in a world with a +2°C global warming relative to pre-industrial levels  
49 (Vautard et al., 2014). Describing the impacts of a +2°C global warming on topics such as water  
50 resources, agriculture or infrastructures is the main aim of the FP7 project IMPACT2C under which this  
51 study was conducted.

52 Future changes in hydrological extremes are still highly uncertain. There is a general consensus that  
53 in most part of the world climate change will result in more rainfall extremes (IPCC, 2012). However, to  
54 what extent this will affect hydrological extremes is still highly uncertain and differs from region to  
55 region. To address this uncertainty many previous studies have used multiple climate models to force a  
56 single hydrological model (e.g. Dankers and Feyen (2009); Feyen et al. (2009)). However recent work,  
57 mainly at global scale, has shown that not only the choice of the climate models affect future change in  
58 hydrological extremes: different hydrological models give sometimes very different results too  
59 (Haddeland et al. (2011); Schewe et al. (2014); Prudhomme et al. (2014); Dankers et al. (2014)). Many  
60 multiple climate and hydrological models impact assessments have used global climate models. These  
61 global climate models tend to have significant biases in the representation of precipitation extremes.  
62 The aim of this study is to improve the assessment of hydrological extreme impacts at the European  
63 continent scale including a better description of uncertainties through the use of multiple hydrological  
64 models and state-of-the art climate projections from the high-resolution CORDEX project. We first focus  
65 on the skill of each hydrological model to simulate extreme floods and hydrological drought (section 3.1)

66 and we next assess the impact that a +2°C warming would have on meteorological variables which are  
67 relevant for flood and drought generation (section 3.2) and on extreme flows (return periods are 10 and  
68 100 years, section 3.3 and 3.4). We finally summarize the results about flood magnitude, droughts  
69 magnitude and drought duration for twelve European cities and with a final assessment on which  
70 European region will be the most affected by extreme flows at +2°C warming.

71

72

## 73 2. Material and data

74

### 75 2.1 Data

76

#### 77 2.1.1 Forcing data

78 This study uses a sub-ensemble from the latest (as of 2014) ensemble of high-resolution,  
79 dynamically downscaled daily climate simulations from CMIP5, 5<sup>th</sup> Coupled Model Intercomparison  
80 Project and CORDEX, A Coordinated Regional Downscaling Experiment (Jacob et al., 2014). The 11  
81 ensemble members were chosen to be representative of the larger ensemble and consist of 5  
82 GCM/RCM combinations and 3 Representative Concentration Pathways (RCPs) (Moss et al. (2010))  
83 (Table 1). One of the main differences with previous impact studies is that rather than focusing on a  
84 future time slice, change is quantified at the 30-year period when each driving GCM reaches +2°C in  
85 global mean temperature relative to pre-industrial levels (1881-1910, Vautard et al. (2014)). Thus, the  
86 ensemble is a representation of a world where greenhouse gas emissions have caused a +2°C global  
87 warming. The climate variables were regridded to a 0.25° resolution grid. As climate models are known  
88 to be affected by some biases that can clearly affect the results (Chen et al., 2013), daily climate

89 variables (precipitation, maximum, minimum and average temperature, dew point temperature,  
90 shortwave and longwave downward radiations) were bias-corrected using quantile mapping (Themeßl  
91 et al. (2011), Themeßl et al. (2012) and Wilcke et al. (2013)). Bias-correction was made using the E-OBS  
92 gridded observational dataset (Haylock et al., 2008) as a reference. These bias corrected data were then  
93 used to run the three hydrological models (1971-2100). Climatic changes were subsequently assessed by  
94 comparing the +2°C period with the baseline period 1971-2000 (Table 1).

### 95 2.1.2 Observed discharge data

96 In order to assess the skills of the hydrological models in representing specific floods and  
97 droughts, we used 428 discharge stations over Europe selected from the Joint Research Center database  
98 which gathers several sources like the Global Runoff Data Centre (2013) data or other publicly available  
99 datasets (e.g., HYDROBanque, CEDEX, Norwegian Water Resources and Energy Directorate, National  
100 river flow archive, Waterinfo.be, eHyd). The original database was filtered to include only stations with  
101 data for the time period 1971-2000 and with no day with missing data in order to not to miss any  
102 extreme discharges. Model performance was validated using the median of the values from each  
103 hydrological model forced by the ensemble of climate models over the 30 years control period.

104

### 105 2.2 Summary of models

106 Three pan-European hydrological models were used to simulate daily discharge: Lisflood (Burek  
107 et al., 2013), E-Hype (Donnelly et al., 2015) and VIC (Liang et al., 1994). The models differ in both  
108 complexity of process description, input data and setup. For example, differences include spatial  
109 resolution (one model is subbasin rather than grid based), description of evapotranspiration and snow  
110 processes, the number of soil layers and depth assumptions and the calibration procedure. A summary  
111 of these models can be found in the Appendix and more details are available in Greuell et al. (2015).

112 Because each of these models has a different output resolution (and in the case of E-HYPE, a high-  
113 resolution subbasin rather than grid output), the output from each model was regridded to a  
114 comparable 0.5°\*0.5° grid.

115

116

## 117 2.3 Methodology for extreme value analysis

118

### 119 2.3.1 High flows

120 We focus here on the discharge magnitude of the 1 in 10 and 1 in 100 year floods (QRP10 and  
121 QRP100). These two return periods were chosen as the first one represents an extreme event occurring  
122 often enough that it is remembered by individuals and communities and the other one is a standard  
123 value used in some countries to design flood protection. In order to compute QRP10 and QRP100 for  
124 each pixel of the grid, we followed a typical extreme value analysis fitting methodology (see e.g. Roudier  
125 and Mahé (2010)). First, the daily maximum discharge was selected for each year of the 30 years long  
126 period, then fitted a Generalized Extreme Value (GEV) distribution using the L-moments and finally, we  
127 calculated QRP10 and QRP100 using the fitted function. The goodness-of-fit was also checked using the  
128 Anderson-Darling test (at 5%), as recommended by Meylan et al. (2008). For each of the 3 hydrological  
129 models, and each of the 11 climate runs, the relative QRP10 and QRP100 change between the +2°C  
130 period and the baseline was computed, resulting in a set of 33 relative changes. To describe this set, in  
131 this paper the median of all the ensemble members is used (the combined climatological and  
132 hydrological model ensemble) rather than the mean, in order to avoid giving excessive weight to  
133 potential outliers. The significance of changes is also assessed using a Wilcox test (5% threshold)  
134 between future period and baseline. Therefore, in all relative change plots, we set as missing values

135 pixel that do not pass the test. Moreover, those that do not pass the goodness-of-fit test were also set  
136 as missing value.

137

138

139

### 140 2.3.2 Low flows

141 We focus for low flows on the same return periods as for the high flow analysis, following the  
142 methodology used by Feyen and Dankers (2009). First the daily discharge time-series is smoothed  
143 applying a seven day moving average in order to remove the day-to-day variations and then for the  
144 magnitudes of low flows ( $Q_{lowRP10}$  and  $Q_{lowRP100}$ ), the lowest smoothed discharge event is selected,  
145 every year. For the duration of low flows a threshold approach is used with the 20<sup>th</sup> flow percentile of  
146 the smoothed flow duration curve as threshold. After computing the 20<sup>th</sup> percentile (for each pixel,  
147 projection and hydrological model) we select all the days that have a smoothed discharge value below  
148 this 20<sup>th</sup> percentile and then we compute the duration of each event below that threshold and select the  
149 maximum drought duration for each year. Finally, for both drought magnitude and duration, the rest of  
150 the methodology is equal to section 2.3.1 (fitting a GEV distribution on each set of 30 values in order to  
151 compute  $Q_{lowRP10}$  and  $Q_{lowRP100}$  magnitude and duration; we also set as missing values pixel that do  
152 not pass the Wilcoxon or Anderson-Darling test)

153

## 154 3. Results and discussion

155

### 156 3.1 Hydrological model validation and selection for ensemble projections



157 A broad and detailed validation of the hydrological models focusing on average conditions is  
158 presented in Greuell et al. (2015). The aim of the validation presented here is to assess the models' skill  
159 to simulate specific indicators used in this study. For each hydrological model we therefore assess if the  
160 median QRP100 (Q<sub>low</sub>RP100) computed based on the 11 bias corrected climate runs is close to the  
161 QRP100 (Q<sub>low</sub>RP100) computed with observed discharge data. Results are shown in Figure S1 (see  
162 supplementary material) for floods and Figure S2 for low flows. The skills of the three hydrological  
163 models are generally better for high than low flows. Lisflood performs slightly better than the other  
164 models, according to the Nash-Sutcliffe Efficiency coefficient (NSE = 0.82) and the Root Mean Squared  
165 Error (RMSE = 961 m<sup>3</sup>/s). However, when focusing on the NSE of the logarithmic modeled and observed  
166 values (NSE(log), used for extreme values, see Krause et al., (2005)), E-Hype is slightly better than  
167 Lisflood. For low flows, Lisflood somewhat overestimates Q<sub>low</sub>RP100 while the other two models  
168 underestimate it. E-Hype has the best performance for low flows (NSE = 0.72, NSE(log)=0.68, RMSE = 75  
169 m<sup>3</sup>/s) compared to Lisflood (NSE=0.54, NSE(log)=0.51, RMSE=95 m<sup>3</sup>/s) and VIC (NSE=0.38,  
170 NSE(log)=0.38, RMSE=112 m<sup>3</sup>/s).

171 A likely explanation for these differences in performance could be the way these models are  
172 tuned to observation data. The Lisflood model is calibrated in individual catchments using a high-  
173 resolution (5 km) interpolated observation data set (EFAS-meteo, see Ntegeka et al. (2013)) and default  
174 parameters in ungauged catchments. The VIC model uses a general set of parameters applicable  
175 anywhere in the model domain, but linked to soil-type and landuse. This parameter tuning was done  
176 using a specific forcing dataset described in Nijssen et al. (2001) .The E-HYPE model also uses a general  
177 set of parameters, linked to soil-type and landuse. For E-HYPE, these parameters were originally  
178 calibrated to a small set of representative gauged basins for each soil and land use using a corrected  
179 ERA-INTERIM forcing data set (see Donnelly et al. (2015)), but for this study an evapotranspiration  
180 parameter was slightly adjusted to better balance the model performance with the E-OBS data set. This

181 was deemed important by the E-HYPE modeling team as the climate scenarios were bias-corrected to  
182 the E- OBS data set and recent studies have shown that bias-correction and climate impact results can  
183 vary between reference observation data sets (e.g. Gutmann et al. (2014)). It is therefore not surprising  
184 that overall E-HYPE performs well when forced with data bias-corrected to E-OBS, at least regarding  
185 mean discharge. Lisflood is likely to perform better in regions where the E-OBS forcing is not significantly  
186 different to the EFAS-meteo data as the other models compromise performance in individual  
187 catchments for relatively good performance in multiple catchments using the same parameters. For  
188 hydrological extremes, again Lisflood can be expected to perform well in calibration catchments. E-HYPE  
189 may outperform VIC for extremes because the representative gauged basins used to calibrate E-HYPE  
190 are generally smaller catchments (1000 to 5000 km<sup>2</sup>) where runoff generating processes dominate over  
191 routing and lake processes.

192           Given the variability in model performance for extremes, and perhaps inability of some models  
193 to successfully represent these extremes, we suggest to remove for the initial selection the hydrological  
194 models that have a NSE and a NSE(log) below 0.5 for the index studied, i.e. QRP100 and QlowRP100.  
195 Based on the NSE threshold, an ensemble consisting of results from E-HYPE, Lisflood, and VIC will be  
196 used for floods, and Lisflood and E-HYPE for low flows.

### 197           3.2 Summary of projected changes in meteorological variables pertaining to floods and droughts

198           We first aim at studying here the future evolution of some of the drivers of low and high flow  
199 changes. We therefore plot in Figure 1 the projected change in maximum annual snowpack (outputs  
200 from the 3 hydrological models) and intense daily rainfall (return period is 10 years, computed using a  
201 GEV distribution using the aforementioned methodology for discharge) at +2°C warming. Results show  
202 that there is generally a clear decrease in maximum snowpack, reflecting the future warming over  
203 Europe. The regions that are less affected like southern Spain or southern Italy are those with currently

204 marginal snowfall. The snowpack decrease is particularly important in Fenno-Scandinavia and the Alps.  
205 Intense rainfall events are projected to increase significantly over the whole continent, with no  
206 particular spatial pattern, which is consistent with previous studies (Madsen et al. (2014); Rajczak et al.  
207 (2013)).

208

### 209 3.3 Floods (QRP10 and QRP100)

210 For changes in flood magnitude there is a clear North to South gradient, with a strong increase  
211 with a strong increase in flood magnitudes south the 60°N line, except for some regions in Bulgaria,  
212 Czech Republic, Poland, the western Balkans, the Baltic countries, and southern Spain where no  
213 significant changes can be detected (Figure 2). Almost everywhere the increase in 100 year floods  
214 (QRP100) is stronger than the 10 year floods (QPR10). Floods are even increasing in areas such as  
215 southern Mediterranean where the average discharge is projected to decrease (Greuell et al., 2015).  
216 However flood changes are consistent with the extreme rainfall changes south of 60°N (Figure 1). Above  
217 the 60°N line, the situation is more heterogeneous with a relatively strong decrease in flood magnitude  
218 in parts of Finland, NW Russia and North of Sweden with the exception of southern Sweden and some  
219 coastal areas in Norway where increases in floods are projected. Projections of decreasing flood  
220 magnitudes are mainly due to the decreases in snowpack in areas where most of the floods are caused  
221 by spring snowmelt in combination with rainfall. Increases in flood magnitude in Scandinavia are mainly  
222 seen in coastal areas where the rain-fed floods will increase (similar results were found by Vormoor et  
223 al. (2015)). Note also that the date of occurrence of annual maximum discharge is expected to be earlier  
224 in spring for these areas (Figure S8) while for the rest of Europe, changes are quite limited.

225 Following this reduction in snow-melt floods in Fenno-Scandinavia and according to the  
226 decreasing snow pattern shown in Figure 1, it is expected a similar decrease in floods would be expected

227 for the Alps. However, our results shown an increase in floods which is also reported in several other  
228 studies (Rojas et al. (2012), Gobiet et al. (2014) or Köplin et al. (2014)). For Switzerland, Köplin et al.  
229 (2014) showed that there are large contrasts in the flood regime in the Alps, and snowmelt-floods only  
230 occur in a limited part of country. Thus, at a  $0.5^{\circ} \times 0.5^{\circ}$  resolution, we are not able to capture some local  
231 decreases of QRP10 and QRP100 due to earlier snowmelt. This hypothesis is confirmed by the same  
232 analysis performed only for Lisflood at  $5 \times 5$  km grid that shows for some areas a decrease of QRP100.  
233 Secondly, we think that at the  $0.5^{\circ} \times 0.5^{\circ}$  resolution over the Alps, there is for most of the pixels a change  
234 of flood regime (from snowmelt to rain-fed or mixed snowmelt-rain-fed) rather than only an earlier  
235 occurrence of snowmelt floods, leading to a different situation compared to Fenno-Scandinavia. This  
236 hypothesis, is supported by Figure S8 which shows (i) a clear reduced number of high flows in June  
237 (southern Alps) and July (rest of the Alps) but (ii) no strong increase in earlier spring. A deeper analysis,  
238 at a finer scale, is needed to fully understand all the changes in the Alps but beyond the scope of the  
239 present paper.

240 Finally, we also provide model-by-model details of these results in the Appendix. Generally, all  
241 three hydrological models show the same pattern. However, in Western Europe and the Mediterranean  
242 area, Lisflood projects a stronger flood increase than the other models.

243 Even if it is difficult to compare with studies that do not focus on exactly the same set of  
244 parameters (climate models, resolution, time period), these results are mainly in line with the recent  
245 literature as reviewed by Madsen et al. (2014) (table 3 in their paper). Out of 22 studies dealing with  
246 future flood projections in Europe, 4 of them are not directly comparable (they differentiate winter and  
247 spring floods), 14 have a global agreement with our findings and 4 give different results. Those four  
248 studies include (i) the UK, where Reynard et al. (2010) find few catchments with changes in flood  
249 frequency above +20% and Kay et al. (2006), with a limited ensemble of driving climate runs, depict a

250 decrease in flood magnitude in south and east England; (ii) France (Seine and Somme rivers) where  
251 Ducharne et al. (2010) cannot conclude on any robust change while we find a significant increase of  
252 flood events and (iii) eastern Germany, Poland and southern Sweden where Rojas et al. (2012) find  
253 decreases in flood magnitude (this latter study which covers the whole continent agrees however with  
254 our results for the rest of Europe). Finally, Dankers et al. (2014) find more contrasted changes for 1 in 30  
255 year flood change in Europe using several GCMs and hydrological models: according to these results,  
256 floods would decrease in large parts of Europe including Greece, Italy and eastern Europe.

257

#### 258 3.4 Low flows: magnitude and duration

259 Low floods ( $Q_{lowRP10}$ ) are expected to decrease for many countries mainly located in the  
260 southern part of Europe: Spain, France, Italy, Greece, the Balkans and also south of the UK and Ireland.  
261 This is mainly due to less rainfall (Rajczak et al., 2013) and also higher potential evapotranspiration in  
262 some regions like Italy (see Figure S4 and Van Vliet et al., 2015). The duration of these droughts is also  
263 increasing (Figure 3, top), especially in Spain. For the rest of Europe, the projections show generally a  
264 decrease of drought magnitude and duration. Moreover, changes are not significant in some areas like  
265 western Germany or southern Sweden: areas with in-significant changes are larger for low flows than  
266 for floods. This reduction of low flow duration and magnitude is mainly caused (i) by less snowfall and  
267 more precipitation for areas with low flows in winter and (ii) by a general increase of rainfall for areas  
268 with low flows in summer (Vautard et al., 2014). Both hydrological models depict generally the same  
269 pattern (see Figure S5 for magnitude). However, for southern Sweden, E-HYPE predicts a small decrease  
270 in low-flow magnitudes and Lisflood an increase. Lisflood also tends to predict a significant increase of  
271  $Q_{lowRP10}$  across Eastern Europe while for E-HYPE this change is also positive, but not significant (white

272 pixels). The same spatial pattern is found for the 100 year return period (see Figure S6) but the results  
273 are generally less significant, see for example southern France.

274 Our results on drought magnitude change are similar to those presented by Forzieri et al. (2014).  
275 Focusing on the Q<sub>lowRP20</sub> they also find for the 2080s a decrease for the Mediterranean and the UK but  
276 their findings for Sweden and Norway show an almost uniform increase of low flows although we  
277 observed a North/South difference that may reflect the different set of climate models used.  
278 Prudhomme et al. (2014), using several climate and hydrological models find a general increase of  
279 hydrological droughts over Europe, but they focus on less extreme droughts, and they use RCP 8.5, at  
280 the end of the century. At a more local scale, other studies agree with our results. For example, in  
281 France, Chauveau et al. (2013) also predict a decrease in low flow magnitude; in Germany Huang et al.  
282 (2014) find a decrease of Q<sub>lowRP50</sub> in 2080 for some areas like the Rhine basin and uncertainty  
283 elsewhere.

### 284 3.5 Uncertainties and summary of the results

285 We analyzed the spread in future change of hydrological extremes using three different  
286 parameters (QRP10, Q<sub>lowRP10</sub> and Q<sub>lowRP10</sub> duration) and focusing on the 1<sup>st</sup> and 3<sup>rd</sup> quartiles of  
287 relative changes distribution (Figure 4 and Figure S7). The spread among results is the largest for  
288 Q<sub>lowRP10</sub> duration, depicted in Figure 4 by the inter-quartile range. Despite this spread, the sign of  
289 Q<sub>lowRP10</sub> duration change (positive or negative) is the same for the first and third quartiles for areas  
290 with a very large spread (e.g. eastern Europe), i.e. the ensemble of projections somewhat agree on the  
291 direction of change, but uncertainty is high. Second, for Q<sub>lowRP10</sub> magnitude the spread between both  
292 quartiles is generally smaller especially in areas like France, the UK, Italy, Portugal and Greece where it is  
293 generally below 20%. However, both quartiles do not have the same sign in all these areas (e.g. the UK  
294 and Italy) thus depicting projections less robust than for Portugal, south of France or south of Spain.

295 Third, for floods (QRP10) the quartiles agreement is slightly better than for QlowRP10 magnitude except  
296 in southern Fenno-scandinavia This general larger uncertainty for low flows is mainly due to (i)  
297 differences between the two models in soil moisture and evapotranspiration calculation which are  
298 directly related to low flows (Greuell et al., 2015) and (ii) the number of selected models is different for  
299 floods (VIC, E-HYPE, Lisflood) and droughts (E-HYPE, Lisflood). Figure S1 and Figure S2 show clearly that  
300 the three hydrological models perform similar for floods while for droughts, Lisflood tends to  
301 overestimate QlowRP10 and E-HYPE to underestimate it, thus resulting in a larger spread. Moreover,  
302 with this kind of assessment using quartiles, the areas with robust results are generally the same than  
303 the ones using the Wilcox test, except for some areas like southern Spain where this latter test gives  
304 significant changes.

305 In order to detect hotspot regions that will be subject to negative changes in several extreme  
306 flow indicators, we combined the floods, drought magnitude and drought duration changes in a single  
307 analysis (Figure 5). In most parts of France, Spain, Portugal, Ireland, Greece and Albania, the projected  
308 changes under +2°C are generally more extreme. We mean by “more extreme” regions where there is a  
309 consistent worsening (of at least 5%) in all the extreme indicators considered for a 10-years return  
310 period: more intense floods (QRP10 >+5%), more intense hydrological droughts (QlowRP10<-5%) and  
311 longer droughts (QlowRP10 duration > +5%). In parts of Norway, Sweden, Finland and western Russia  
312 future warming will see a reduction in both streamflow floods and droughts.

313

#### 314 4. Conclusion

315 Our aim is to make a robust assessment of the impact that a +2°C global warming would have on  
316 hydrological extremes (floods, droughts magnitude and duration) in Europe by using an ensemble of  
317 eleven high-resolution RCM outputs and three pan-European hydrological models. Results show that

318 such a warming could increase flood magnitudes (10 years and 100 years return period) significantly in  
319 most parts of Europe (e.g. about +20% close to London; and Warsaw for QRP10), even for areas where  
320 the annual rainfall is expected to decrease in the future, e.g. Spain (about +10% QRP10 close to Sevilla).  
321 However, in Fenno-Scandinavia the situation is more contrasted with (i) a large area that is expected to  
322 have less intense snowmelt floods, occurring earlier in spring except and (ii) the southern and coastal  
323 areas of Fenno-Scandinavia where we predict an increase of rain-fed flood magnitude. In the Alps, even  
324 though snowpack is also projected to reduce, floods are expected to increase generally due to a change  
325 of flood regime from snowmelt to rain-fed and possibly because the spatial resolution of this study  
326 (0.5°\*0.5°) potentially hides local decreases of intense snowmelt floods. Moreover, despite some  
327 significant differences among the hydrological models' structure and calibration, the results are quite  
328 similar from one model to another and consistent with other studies.

329 Future changes in hydrological drought magnitude and duration show contrasting patterns across  
330 Europe. Our projections show that for large areas of Italy, France, Spain, Greece, the Balkans, Ireland  
331 and the UK, droughts will become more intense and longer mainly due to less rainfall and higher  
332 evapotranspiration, in some areas. The sign of these changes is particularly robust in southern France,  
333 parts of Spain, Portugal and Greece. For the rest of Europe changes in droughts are not significant or  
334 there is a reduction of droughts length and magnitude, especially in northern Fenno-Scandinavia and  
335 Western Russia where the sign of the changes is also very robust.

336 Our results show that for a significant part of Europe there will be a clear intensification of the  
337 hydrological cycle resulting in both increases in droughts and floods. Extreme flows will be particularly  
338 harmful in Spain, Greece, France, Ireland and Albania: it is thus urgent to integrate these future changes  
339 for policy making in water resources management and flood protection design.

340



341 **Acknowledgments**

342 The authors would like to thank the FP7 project IMPACT2C and all the contributing members for funding  
343 this study and providing climate data. Moreover, we thank Goncalo Gomes from JRC for his help with  
344 observed discharges and Alessandra Bianchi for GIS support. We finally thank three anonymous  
345 reviewers for their helpful comments.

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486 **Tables**

487

RCM	Driving GCM	RCP	+2°C period
CSC-REMO2009	MPI-ESM-LR	8.5	2030-2059
		4.5	2050-2079
		2.6	2071-2100
SMHI-RCA4	HadGEM2-ES	8.5	2016-2045
		4.5	2023-2053
SMHI-RCA4	EC-EARTH	8.5	2027-2056
		4.5	2042-2071
		2.6	2071-2100
KNMI-RACMO22E	EC-EARTH	8.5	2028-2057
		4.5	2042-2071
IPSL-WRF331F	IPSL-CM5A-MR	4.5	2028-2057

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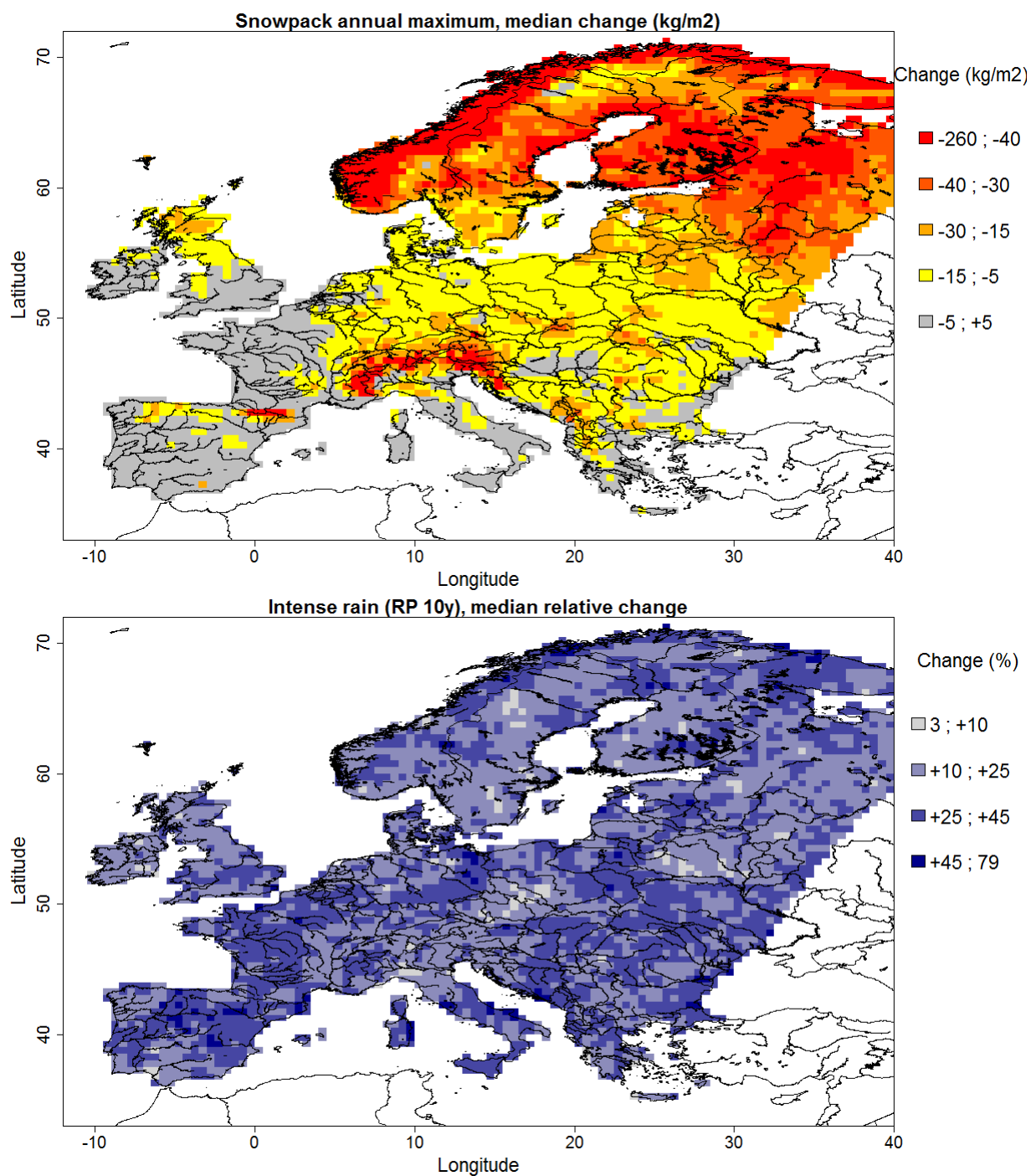
**Table 1: Summary of the 11 climate projections used in this study (RCP, GCM, RCM and +2°C period)**

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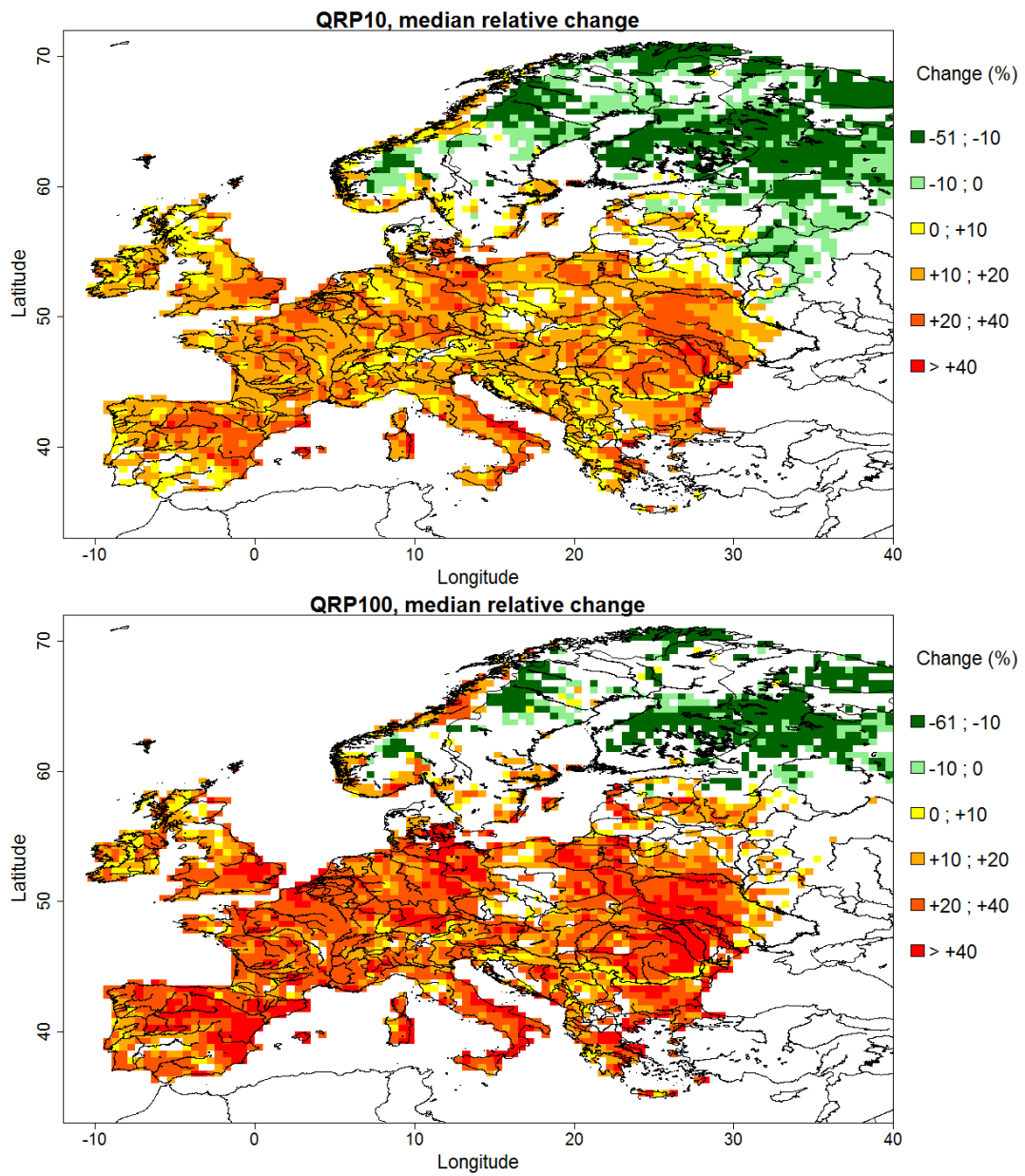
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495 Figure 1: Change in maximum annual snowpack (all three hydrological models, top) and change in intense rain (RP10, all  
496 three hydrological models, bottom)





498

499 **Figure 2: high flows median relative change, for two different return periods; RP10 (top), and RP100 (bottom). The median is**  
 500 **computed over 33 members. Only significant changes (i.e. passing the Wilcox test at 5%) are shown here.**

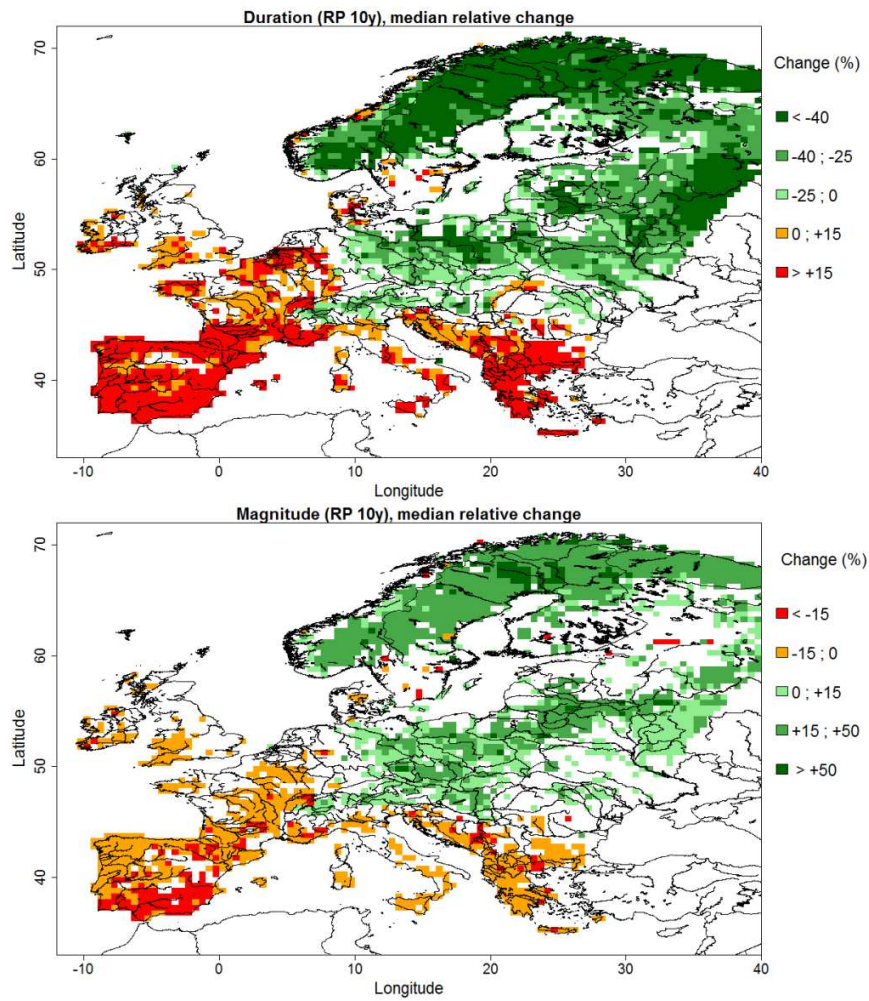
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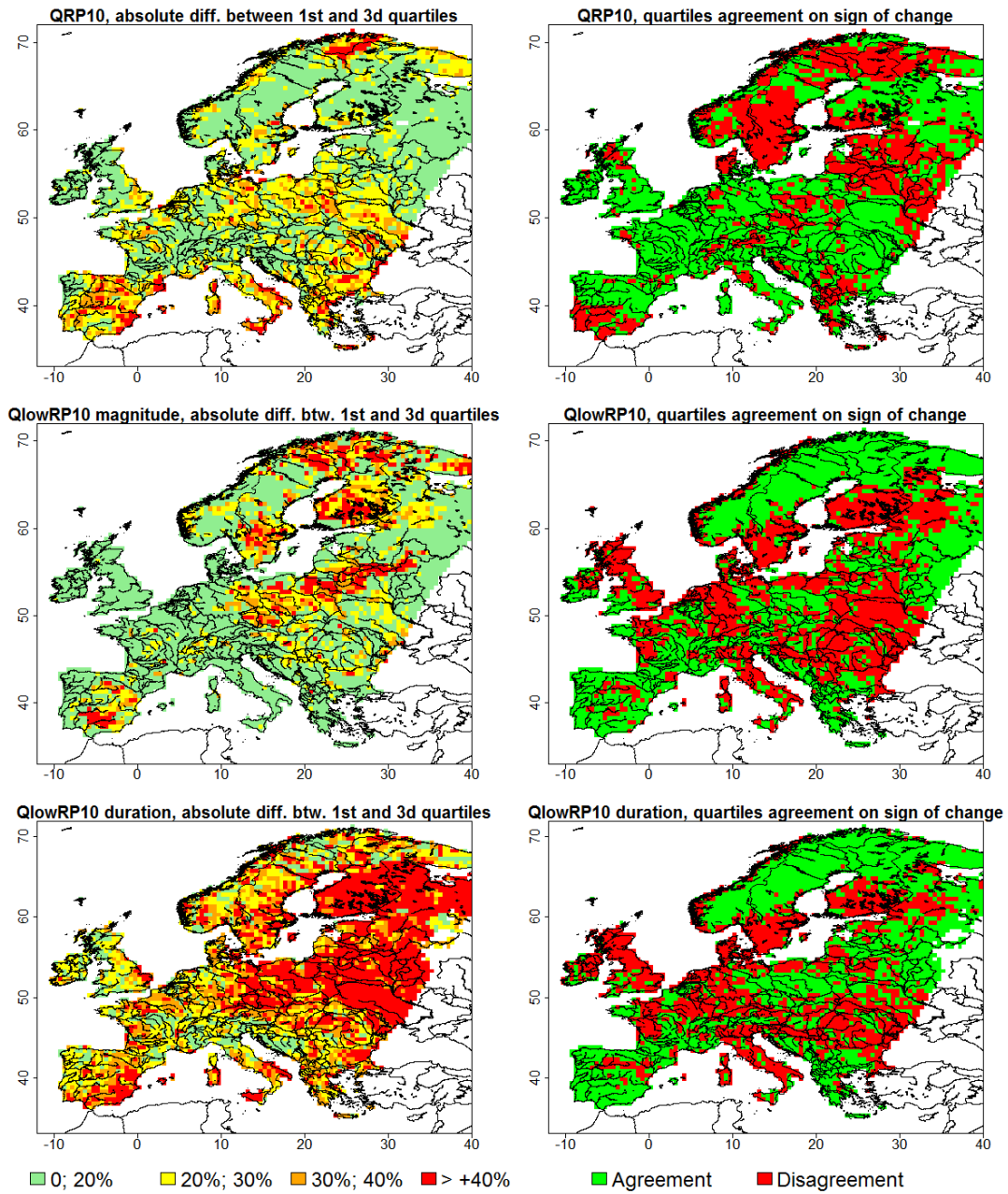
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507 **Figure 3: characteristics of low flows (RP10): duration (top) and magnitude (bottom). The median is computed over 22**  
508 **ensemble members. Only significant changes (i.e. passing the Wilcox test at 5%) are shown here. When  $Q_{lowRP10}$  is zero for**  
509 **the baseline period, we set the relative change as missing value.**

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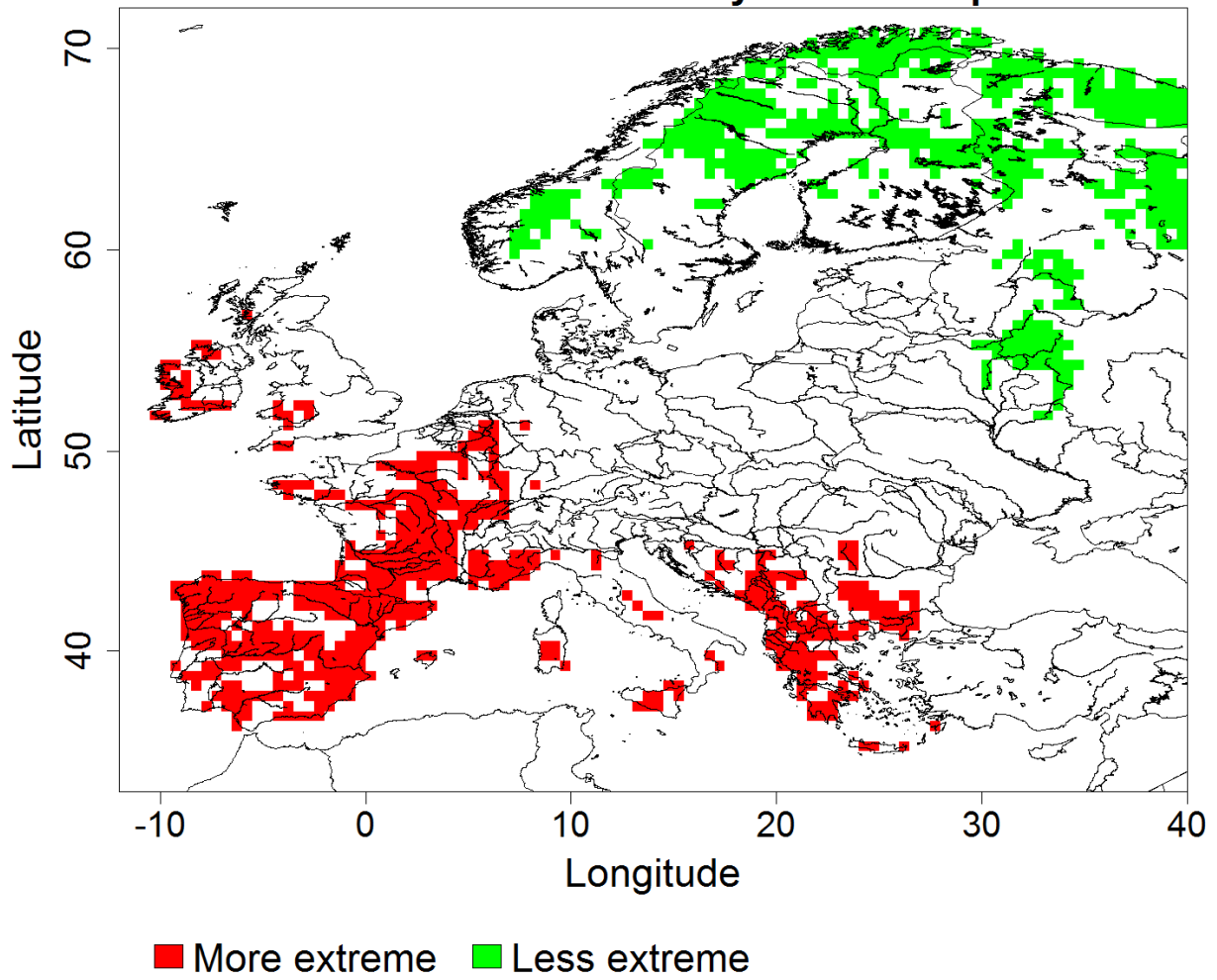
512 **Figure 4: quartiles of relative change. Left column: absolute difference between the 1<sup>st</sup> and 3<sup>rd</sup> quartiles (in**

513 **%) for QRP10 (top), QlowRP10 magnitude (middle) and QlowRP10 duration (bottom). Right column: agreement on the sign**

514 **of the 1<sup>st</sup> and 3<sup>rd</sup> quartiles (e.g. the green area means that both quartiles are positive or negative)**

515

### General assessment for 10-years return period



516

517 Figure 5: summary of the impacts of extreme discharge (return period is 10 years) under a +2C warming. Green area means  
518 that (i) QRP10 change < -5%, (ii) QRPlow10 change > +5% and (iii) QRPlow10 duration change < -5%. We show here only pixels  
519 where all three change are statistically significant.

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## Supplementary material

### 525 **A1. Summary of models**

#### 526 • **E-HYPE**

527         The E-HYPE model (Donnelly et al., 2015) is a pan-European application of the Hydrological  
528 Predictions for the Environment (HYPE, Lindström et al. (2010)) which simulates hydrological variables in  
529 more than 35000 sub-basins (median size 200 km<sup>2</sup>) across Europe. A modified Hargreaves-Samani  
530 equation using daily minimum and maximum temperature and a time constant radiation that is  
531 assumed to vary with latitude is used to compute evapotranspiration. Snow melt is calculated using a  
532 degree day method. In HYPE evapotranspiration, snow and runoff generation calculations are made for  
533 hydrological response units (HRUs) representing unique land use and soil type combinations and  
534 consisting of up to three soil layers down to 2.5 m. Each sub-basin may consist of any number of HRUs.  
535 The model is forced by daily precipitation and temperature and then calculates flow paths in the soil  
536 based on snow melt, evapotranspiration, surface runoff, infiltration, percolation, macropore flow, tile  
537 drainage, and lateral outflow to the stream from soil layers with water content above field capacity.  
538 HRUs are connected directly to the stream and act in parallel. The groundwater level in each HRU is  
539 fluctuating, may saturate the soil layers and water may percolate between sub-basins. Routing of the  
540 runoff is made along local and main rivers within each sub-basin and includes delay and dampening  
541 processes. Lakes, where they exist, cause delay of flow using the weir equation and may be local (off the  
542 main river) or main (on the main river).

543

544

545       •   **LISFLOOD**

546           LISFLOOD is a GIS-based spatially-distributed hydrological rainfall-runoff model, which includes a  
547 one-dimensional hydrodynamic channel routing mode. Driven by meteorological forcing data  
548 (precipitation, temperature, potential evapotranspiration, and evaporation rates for open water and  
549 bare soil surfaces), LISFLOOD calculates a complete water balance at every daily time step and every grid  
550 cell defined in the modelled domain (current resolution is 5\*5km). Basically, the model is made up of  
551 the following components: (i) a 2-layer soil water balance sub-model, (ii) sub-models for the simulation  
552 of groundwater and subsurface flow (using 2 parallel interconnected linear reservoirs), (iii) a sub-model  
553 for the routing of surface runoff to the nearest river channel, (iv) a sub-model for the routing of channel  
554 flow. The processes that are simulated by the model include snow melt, infiltration, interception of  
555 rainfall, leaf drainage, evaporation and water uptake by vegetation, surface runoff, preferential flow  
556 (bypass of soil layer), exchange of soil moisture between the two soil layers and drainage to the  
557 groundwater, sub-surface and groundwater flow, and flow through river channels. Runoff produced for  
558 every grid cell is routed through the river network using a kinematic wave approach.

559

560       •   **VIC**

561           The Variable Infiltration Capacity model (VIC) performed calculations on a regular 0.5x0.5  
562 degrees lat-lon grid. In VIC each grid cell is subdivided into an arbitrary number of tiles, each covered by  
563 a particular vegetation type. Evapotranspiration is computed for each tile, whereas the soil is uniform  
564 within each grid cell. The soil consists of 3 layers with a total depth of ~3 m for most cells. The energy  
565 balance approach of Penman-Monteith (Shuttleworth, 1993) is used to compute evapotranspiration,  
566 requiring a forcing consisting of atmospheric temperature and humidity, wind speed and incoming  
567 radiation. Penman-Monteith is also employed to simulate snow melt. VIC owes its name to the

568 treatment of surface runoff, which is generated from net precipitation under the assumption that the  
569 infiltration capacity varies within each grid cell according to a one-parameter distribution function  
570 (Wood et al., 1992). Routing of the runoff is taken into account with the model described in Lohmann et  
571 al. (1996). This model first transports the runoff generated in a grid cell to the main river leaving the grid  
572 cell with a unit hydrograph and then routes the water through the river network connecting the cells.

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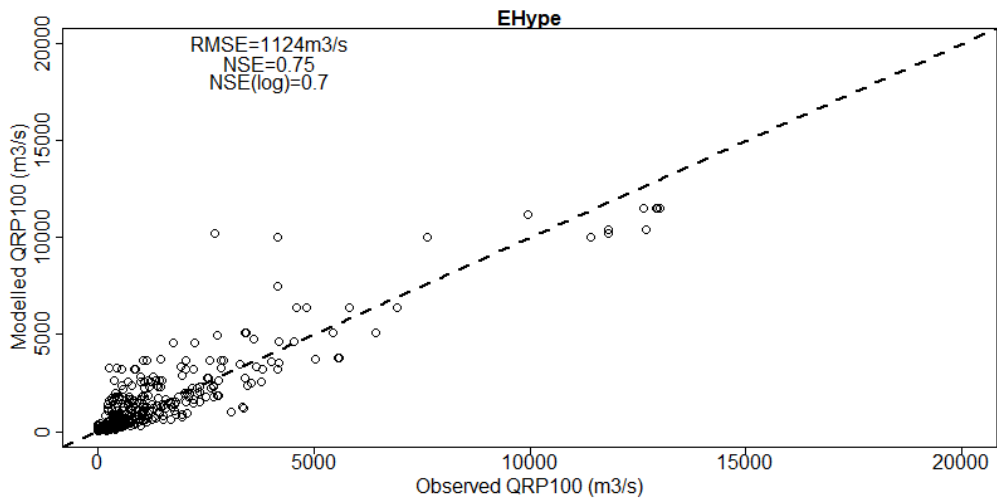
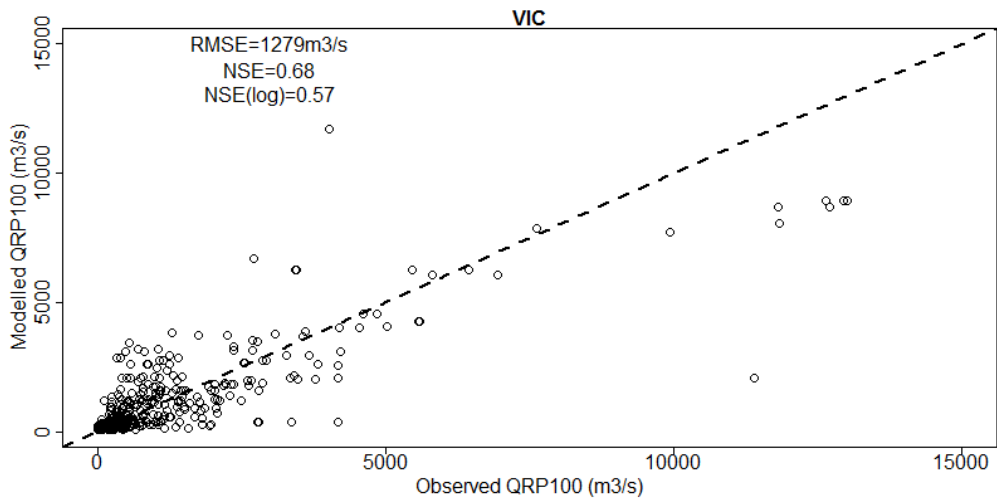
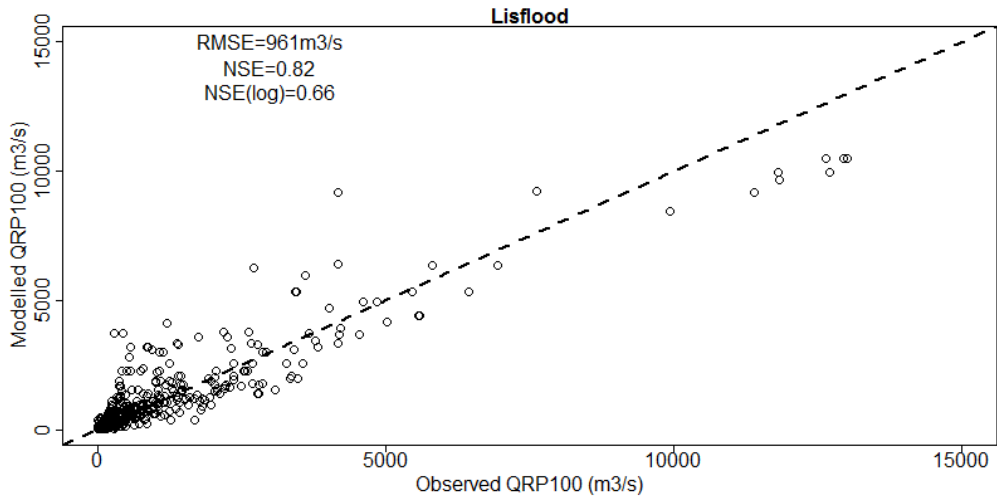
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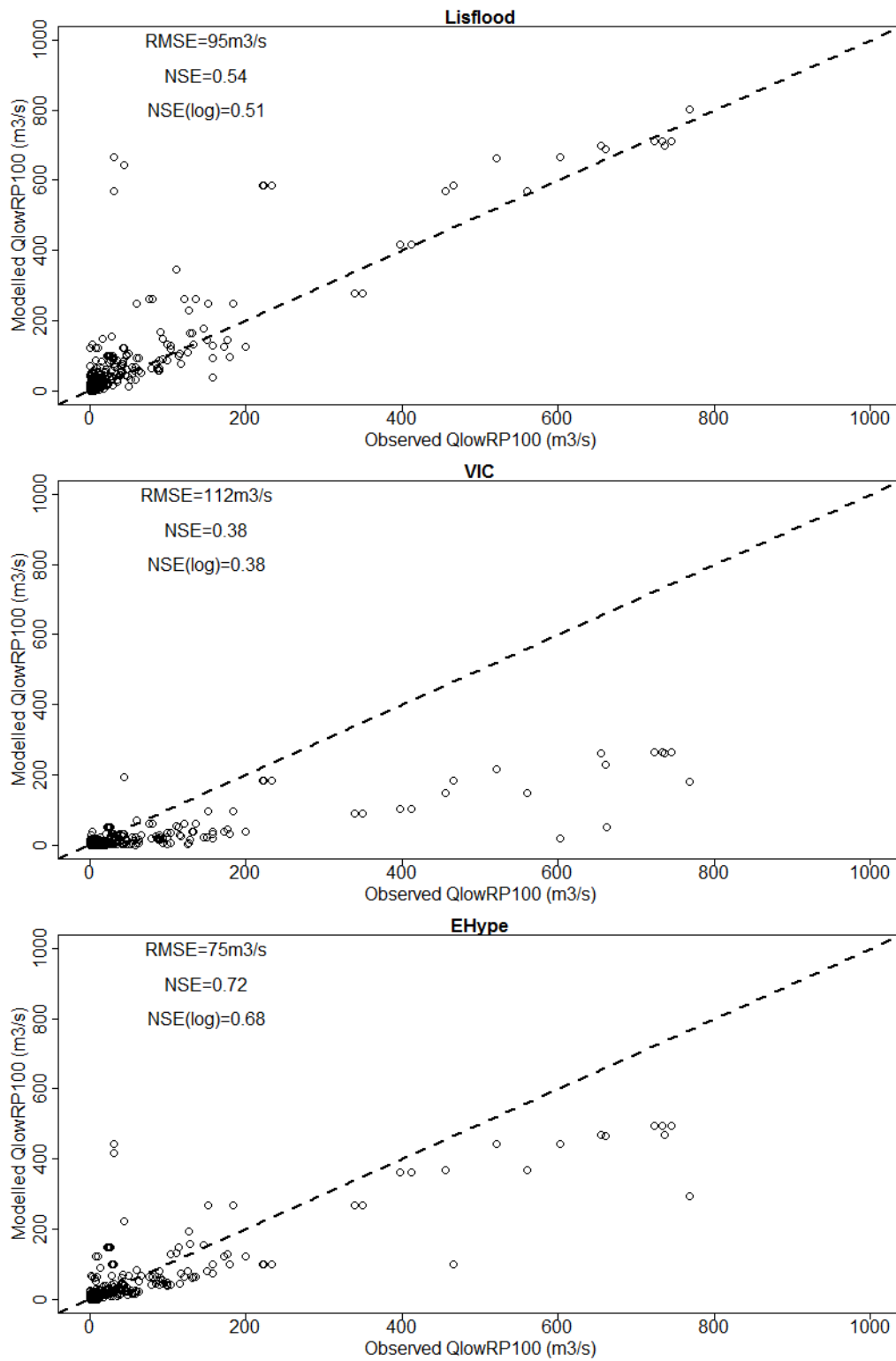
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586 A2. Validation





588 Figure S1: Observed QRP100 428 stations, years 1971/2000) vs. modelled one (median QRP100 over 5 climate members), for  
589 each of the three hydrological models. The dashed line is  $y=x$ , NSE is Nash-Sutcliff Efficiency coefficient, NSE(log) is Nash-  
590 Sutcliff Efficiency coefficient of  $\log(\text{observed QRP100})$  vs.  $\log(\text{modeled QRP100})$  and RMSE is the root mean squared error.

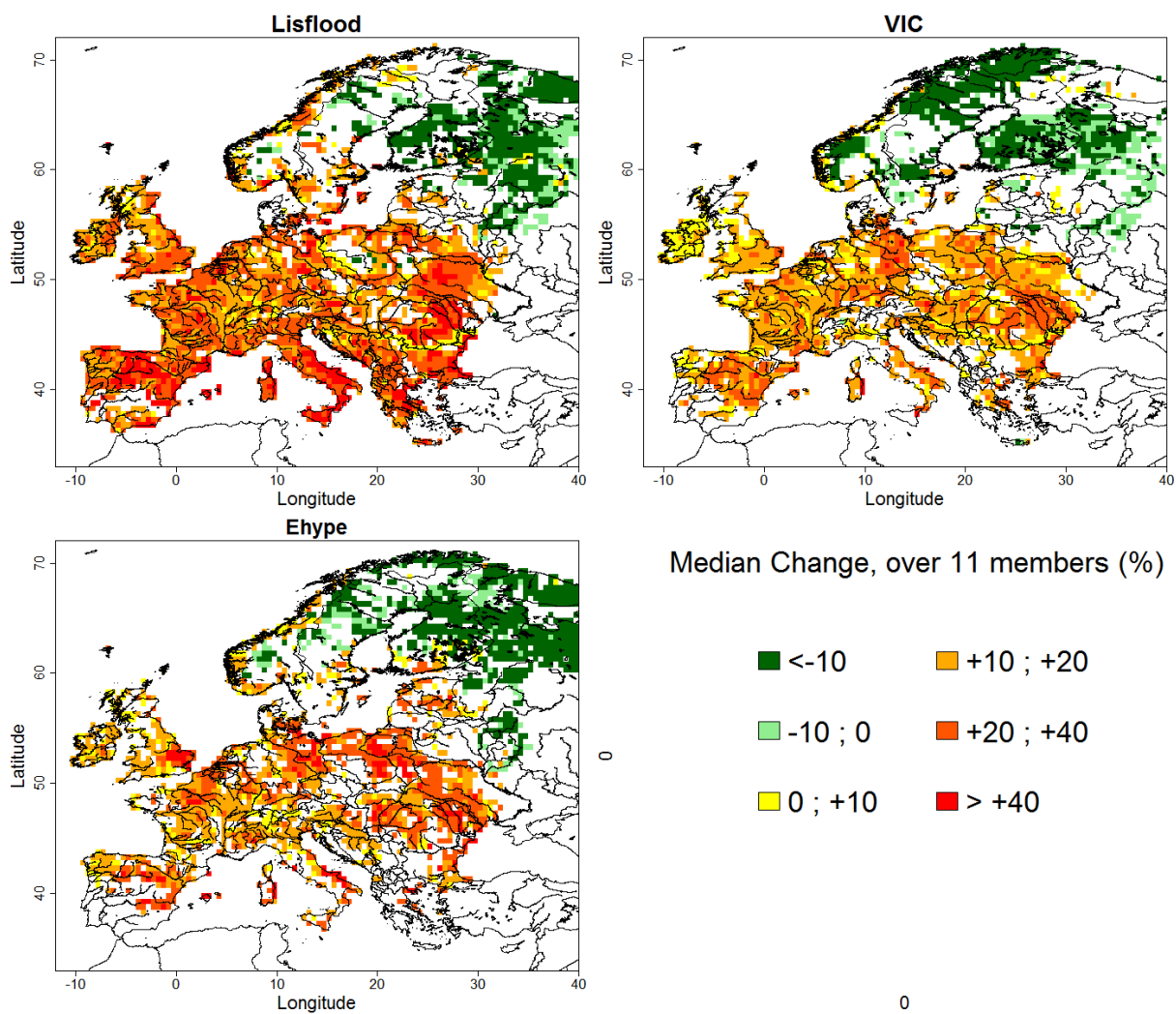


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Figure S2: same as Fig 1 but for QlowRP100

593 A3. Floods, model by model, QRP10



594

595 **Figure S3: QRP10 median change for each of the 3 hydrological models. Median is computed over 11 members for each plot.**

596 **Only significant changes are shown here.**

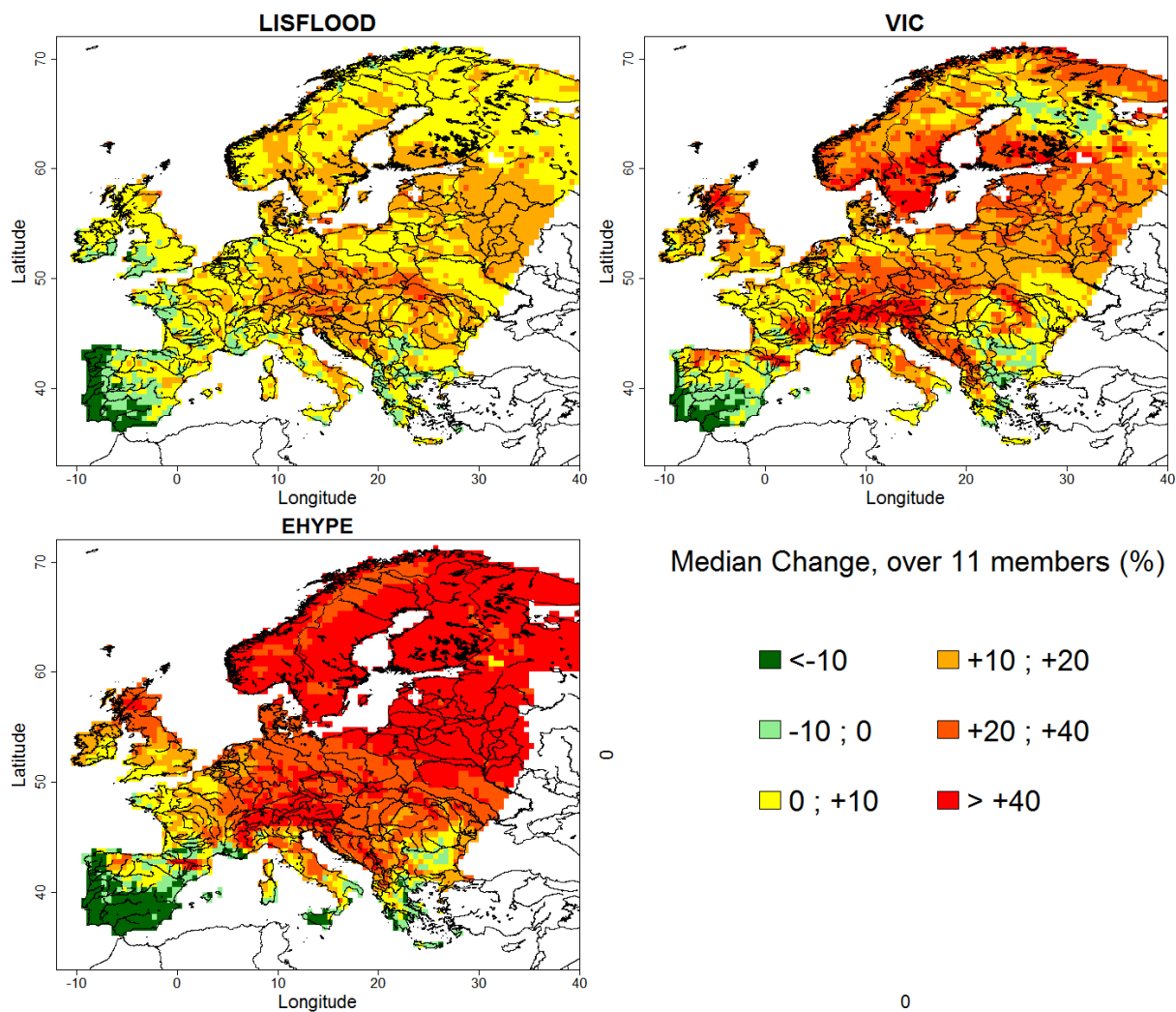
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601 A4. Evapotranspiration change, model by model

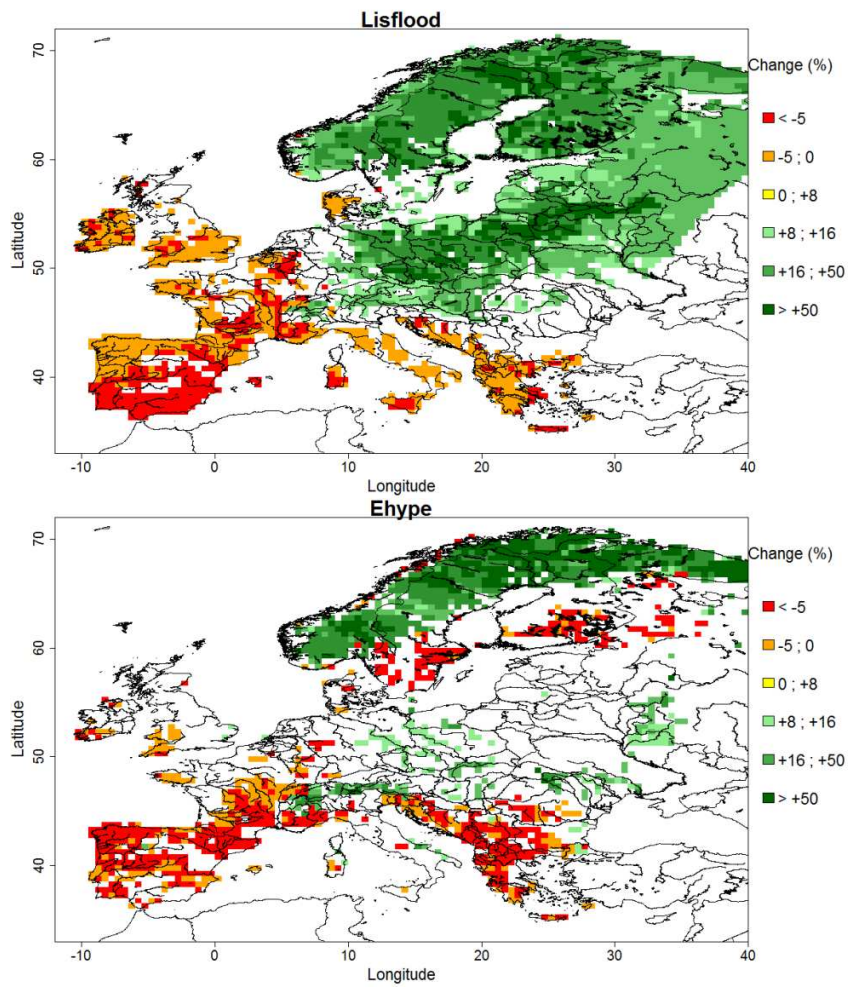


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603 **Figure S 4: evapotranspiration relative change (%), for each of the three models. All pixels are shown.**

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605 A5. Droughts magnitude, model by model



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607 **Figure S5: QlowRP10 magnitude relative change, for both hydrological models used for low flows: Lisflood (top) and E-HYPE**  
 608 **(bottom). The median is computed over 11 members. Only significant changes are shown here. When QlowRP10 is zero for**  
 609 **the baseline period, we set the relative change as missing value**

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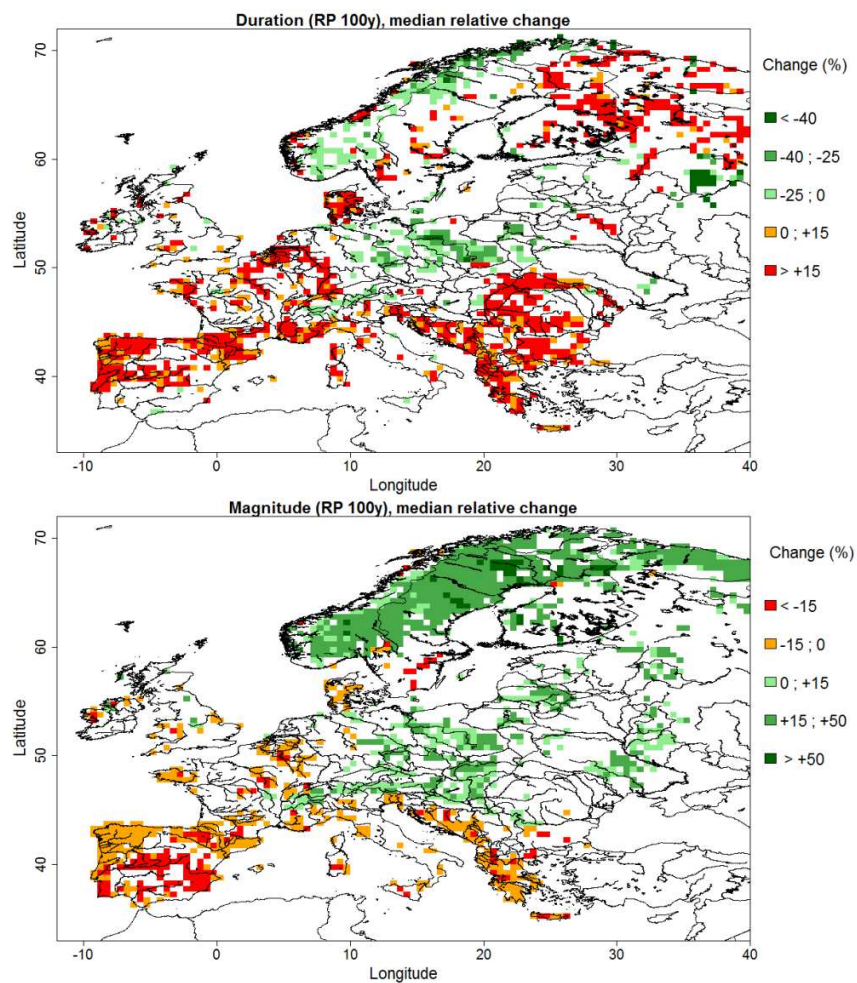
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615 A6. Drought magnitude and duration, RP100



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617 **Figure S 6: characteristics of low flows (RP100): duration (top) and magnitude (bottom). The median is computed over 22**  
618 **ensemble members. Only significant changes are shown here. When QlowRP100 is zero for the baseline period, we set the**  
619 **relative change as missing value.**

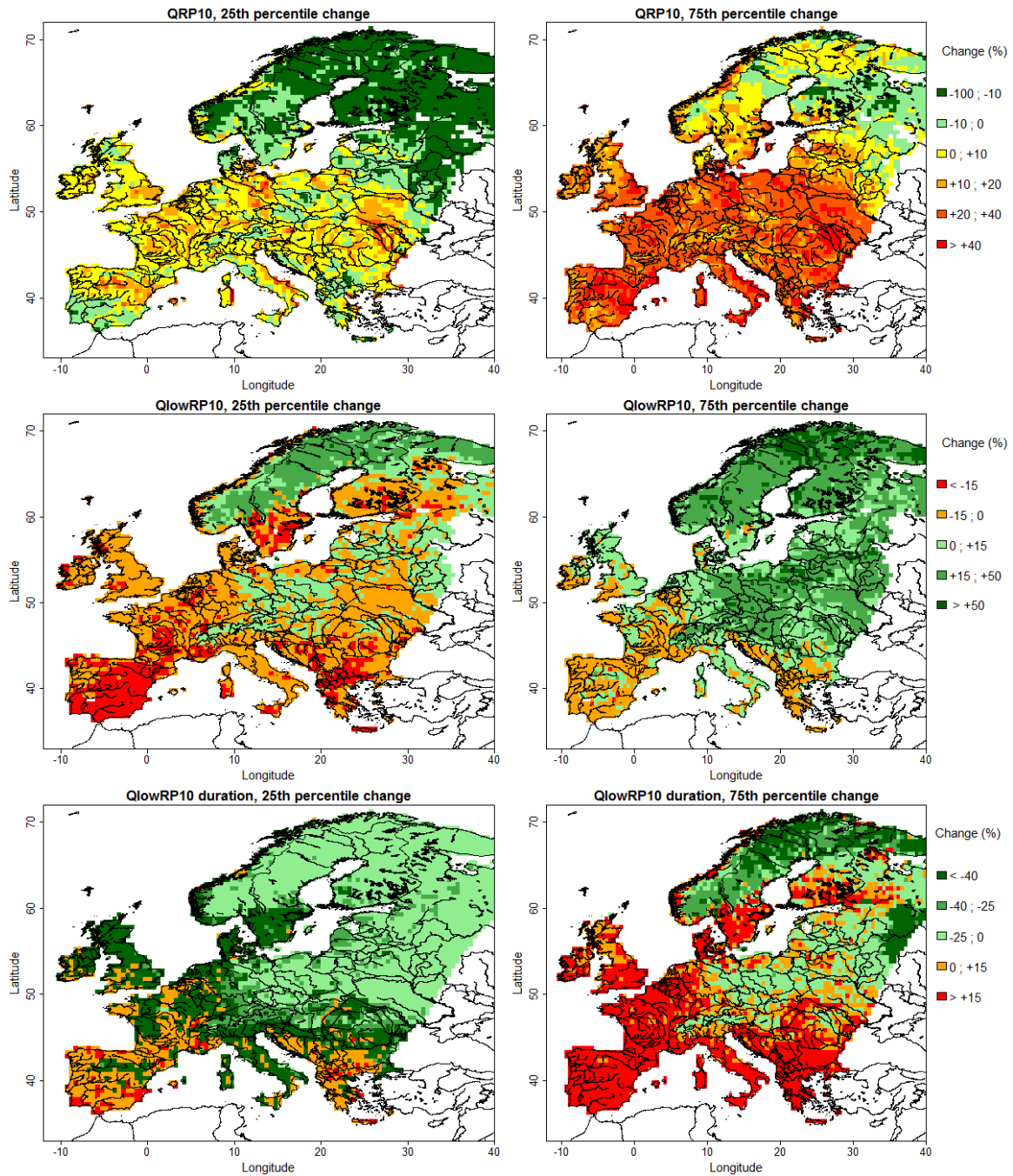
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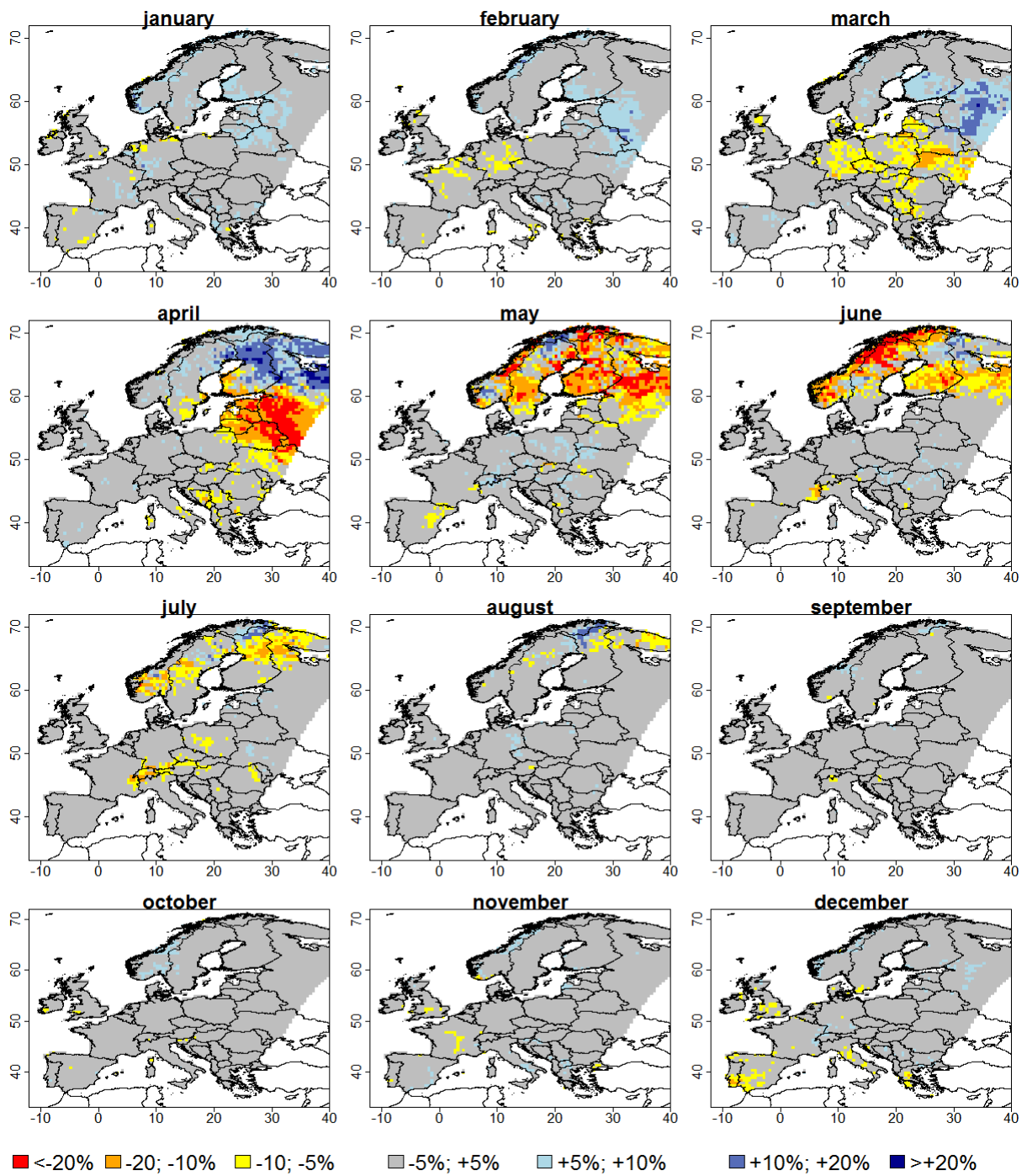
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624 A7. 25<sup>th</sup> percentile and 75<sup>th</sup> percentile change



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626 Figure S7: 25<sup>th</sup> percentile (left column) and 75<sup>th</sup> percentile (right one) relative change for QRP10 (top row), QlowRP10 (middle  
627 row) and QlowRP10 duration (bottom row). For each pixel, the percentiles are computed over 33 members (QRP10) or 22  
628 members (QlowRP10 and QlowRP10 duration).



629

630 **Figure S8: relative change (% , between +2C period and baseline) in occurrence of maximum annual discharge, month by**  
 631 **month. Blue areas mean that in the future, according to the 33 members, the annual maximum discharge occurs more**  
 632 **frequently during the specific month.**



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