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On the benefit of high-resolution climate simulations in impact studies of hydrological extremes

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Abstract

We investigated the effect of changing the horizontal resolution of a regional climate model (RCM) on the simulation of hydrological extremes. We employed the results of three experiments of the RCM HIRHAM using a grid size of approximately 12, 25 and 50 km. These simulations were used to drive the hydrological model LISFLOOD, developed for flood forecasting at European scale. The discharge simulations of LISFLOOD were compared with statistics of observed river runoff at 209 gauging stations across Europe. The largest discrepancies in peak flow occurred in climates with a seasonal snow cover, which may be explained by inaccuracies in the simulated precipitation that accumulate over winter. Although previous studies have found that high resolution climate simulations result in more realistic patterns of extreme precipitation, especially in mountainous regions, we did not find conclusive evidence that the 12-km HIRHAM run generally yields a better simulation of peak discharges. At some gauging stations the model performance is increasing with increasing horizontal resolution of the RCM, while at other stations it is decreasing. However, the differences between the three experiments become less important in larger river basins. Above about 30 000 km^2 and 120 000 km^2, respectively, the 25- and 50-km runs generally provided a good approximation of the simulations based on the 12-km climatology. Under the A2 scenario of climate change, the changes in extreme discharge levels were similar between the three experiments at continental scale. At the scale of individual river basins, however, there were occasionally important differences. If we assume the 12-km HIRHAM simulation to be more realistic, the use of lower-resolution climate simulations may lead to an underestimation of future flood hazard. This means that results obtained with lower-resolution RCM simulations should be interpreted with care, as the grid scale of the climate model adds to the uncertainty.

1 Introduction

In the past two decades there have been a growing number of studies on the potential impact of anthropogenic climate change on hydrology and water resources. Early studies, such as that by Gleick (1987), resorted to relatively crude methods to obtain scenarios of climate change. As a direct use of climate model output would lead to unacceptable bias in the results, the common approach was to perturb a historical time series of meteorological observations with a climate change signal. This perturbed series was then used to drive a hydrological model (Leavesly, 1994). Estimates of climate change were often based on the difference between simulations of current and future climate by one or more General Circulation Models (GCMs). The motivation behind this so-called "delta-change" approach was that the GCMs were believed to simulate relative changes more reliably than the absolute climate conditions themselves (Hay et al., 2000).

Present-day Atmosphere-Ocean General Circulation Models (AOGCMs) realistically reproduce most features of the Earth's climate at global and continental scales, including some of the most important modes of climate variability and the global statistics of extreme events (Randall et al., 2008). Nevertheless, the models still show significant errors, especially at smaller scales. Current AOGCMs typically run at a horizontal resolution of 400 to 125 km but cannot resolve the mesoscale, sub-grid processes that determine the hydrological response of a river catchment. This is particularly true when looking at hydrometeorological extremes like floods, as the coarse resolution of these global models does not allow a realistic simulation of synoptic or regionally induced circulation (Christensen and Christensen, 2004). As a result, many AOGCMs tend to underestimate both the frequency and tendency of heavy precipitation events (Sun et al., 2006).

Over the years, two main approaches have been developed to "downscale" the information at the AOGCM grid to smaller scales: dynamical and statistical downscaling (Christensen et al., 2007). The latter relies on empirical relationships based on observed data, and applies these on climate model output. Although computationally inexpensive, they require sufficient observational data for a long enough period, which limits its application to larger areas. Moreover, the implicit assumption that the statistical relationships also hold under a different climate forcing cannot be verified, and feedback processes are not taken into account. In dynamical downscaling, high-resolution regional climate models (RCMs) are run over a limited area, with boundary conditions coming from either observation-based datasets or lower-resolution AOGCM simulations. In this way, a coherent simulation of multiple climate variables based on physical principles is obtained, and nonlinear effects are potentially taken into account. RCMs have been proven to realistically reproduce a broad range of climates around the world (Christensen et al., 2007).

In recent years the horizontal resolution of RCM experiments has increased considerably. In the European PRUDENCE project (Prediction of Regional scenarios and Uncertainties for Defining EuropeaN Climate change risks and Effects, 2002–2005) (Christensen et al., 2007), eight different RCMs produced climate simulations at a grid scale of approximately 50 km. In its follow-up project ENSEMBLES (Hewitt, 2005) that started in 2004, RCM simulations are performed at about 25 km resolution. Still higher resolution experiments using a grid of approximately 18–12 km have been done in a number of studies (e.g. Christensen et al., 1998; Kleinn et al., 2005; Kotlarski et al., 2005) albeit mostly for smaller areas or a limited number of years.

This advancement in modelling resolution raises the question whether such high resolution simulations provide a better representation of the climate, and particularly its extremes, for use in hydrological impact studies, and whether they result in a substantially different climate signal. In other words, what is the added value of very high resolution RCM experiments when simulating changes in hydrological extremes, such as river floods? For the Upper Danube Basin, Dankers et al. (2007) compared two simulations of the regional model HIRHAM, one with a grid scale of 50 km and one of 12 km, and found that the higher resolution experiment gave a much more realistic representation of the orographic patterns in extreme precipitation in this area. Yet, when the simulations were used to drive a hydrological model to assess flood hazards, the difference between the two RCM experiments turned out to be small, especially in the larger catchments.

The present paper aims at answering the above question by using three different realizations of the HIRHAM model for both current and future climate conditions in Europe, with a grid scale of about 50, 25 and 12 km, respectively. These simulations were used to drive a large-scale hydrological model of the European continent that has been developed for operational flood forecasting. This set-up builds on the approach of Dankers and Feyen (2008), but here we focus on the differences that are induced by the changes in resolution of the climate model.

In the following section we provide a short description of the models that have been used in this study. In Sect. 3, we compare the three simulations for the present climate with observations of river discharge at more than 200 gauging stations in Europe, and with each other. The differences in climate change signal are explored in Sect. 4, followed by discussion and conclusions at the end.

2 Methodology

2.1 Climate model

We used three experiments of the regional climate model HIRHAM of the Danish Meteorological Institute (Christensen et al., 1996) that were performed within the framework of the PRUDENCE project (Christensen et al., 2007). In each of these experiments the boundary conditions were derived from the HadAM3H high-resolution global atmosphere model of the Hadley Centre in the UK (Pope et al., 2000). Each of these experiments consisted of two 30-year time slices: a control run with a greenhouse gas forcing corresponding to 1961–1990, and a scenario run corresponding to 2071–2100 with greenhouse forcing according to the A2 scenario of the Intergovernmental Panel on Climate Change (IPCC) (Nakicenovic et al., 2000). The prime difference be-

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tween the experiments was the horizontal resolution of the HIRHAM model, namely 0.44° (∼50 km), 0.22° (∼25 km) and 0.11° (∼12 km). These experiments are hereafter referred to as H50CL (H50A2), H25CL (H25A2) and H12CL (H12A2), respectively, where the last two letters denote the control (CL) and scenario (A2) runs. Furthermore, it should be noted that in the scenario period the 12-km simulation used sea surface temperatures from a different RCM over the Baltic Sea, as the driving GCM shows a large warming in this area. A more detailed description and analysis of these HIRHAM runs is given in Christensen and Christensen (2007). The performance of HIRHAM in simulating extreme precipitation has been evaluated in different areas in a number of recent studies, e.g. by Frei et al. (2006), Beniston et al. (2007), Dankers et al. (2007), Fowler et al. (2007) and, at European scale, by May (2007).

2.2 Hydrological model

As in Dankers and Feyen (2008), the HIRHAM simulations of temperature, precipitation, solar and thermal radiation, humidity and wind speed were used to drive the hydrological model LISFLOOD (De Roo et al., 2000; Van der Knijff et al., 2008). This model has been developed for operational flood forecasting at the European scale and is a combination of a grid-based water balance model and a one-dimensional hydrodynamic channel flow routing model. The current set-up uses a 5-km grid and has been calibrated against discharge records in 231 catchments and subcatchments in Europe (see Dankers and Feyen (2008) and Van der Knijff et al. (2008) for more details). The HIRHAM output was regridded to the 5-km grid scale of LISFLOOD without any further downscaling or altitude correction. This means that any bias in especially the precipitation fields will directly influence the LISFLOOD simulations. However, our main objective in the present study is to compare the differences in the simulation of discharge extremes induced by the different horizontal resolution of the climate model experiments.

3 Present climate validation

3.1 Comparison with observations

We compared the river discharge simulated by LISFLOOD when being driven by the three HIRHAM control experiments with observations at 209 gauging stations across Europe for which at least 30 years of discharge data were available. The distribution of these stations is very uneven, as hardly any data were available for the Mediterranean countries (see Fig. 1). The observations cover either the period 1960–1990 or 1970–2000. It should be noted that the HadAM3H and HIRHAM simulations for the control period (i.e. 1961–1990) do not reproduce the historical weather of this period. As Boer and Lambert (2001) point out, the aim of a climate simulation is not to predict the actual evolution of the climate system because the internally generated natural variability is essentially unpredictable. A simulation should rather be seen as one of many possible evolutions, given the forcing of the system. The skill of a climate simulation is in correctly reproducing the forced deterministic components, such as the mean, annual and diurnal cycles or greenhouse-induced climate change, together with pertinent statistics of the natural variability (Boer and Lambert, 2001). The implication is that a direct, day-to-day or even year-to-year comparison of simulated discharges with observations is senseless. Since the evolution of the weather or natural variability 'noise' is different in the simulation, such a comparison is bound to break down. Instead, one should pay attention to what Boer and Lambert (2001) call the "second order statistics", such as the mean square difference, which can tell how well the simulation is reproducing the deterministic components of the climate system.

Since our focus is on the simulation of peak flows, we compared the annual maximum discharges (Q_{max}) in both the observed and simulated time series at the 209 gauging stations. After ranking the 30 Q_{max} values from high to low we calculated the normalised root mean square error (*nRMSE*), which is the *RMSE* divided by the mean of the observed values:

$$nRMSE = \frac{1}{\overline{Q_{\max,o}}} \times \frac{\sqrt{\sum_{i=1}^{n} \left(Q_{\max,s}(i) - Q_{\max,o}(i)\right)^2}}{n} \quad (1)$$

where $Q_{\max,o}$ and $Q_{\max,s}$ denote the $Q_{\max}$ values of the observed and simulated discharges, respectively, $\overline{Q_{\max,o}}$ is the mean of the observed maximum values, i is the order after ranking on magnitude, and n is the number of years. Note that the series of (in this case) 30 maximum values usually provides the basis for fitting a generalised extreme value (GEV) distribution under the assumption of being independent and identically distributed observations (see Katz et al., 2002). By focusing on the $Q_{\max}$ values, however, we preclude any differences arising from fitting an extreme value distribution to the data.

In Fig. 1 the *nRMSE* is plotted for all 209 stations, whereby the colour of the symbols indicates which of the three HIRHAM experiments gives the smallest error. Note that the only difference between the three experiments is the horizontal resolution of the climate model, since the hydrological model setup is the same in all model runs. To be fair, any differences between simulated and observed discharges are not only due to bias in the RCM climatology, but also to errors and uncertainties in LISFLOOD. In particular, the current model set-up does not take into account river regulation, while almost all rivers included in the comparison are regulated to at least some extent (see Dynesius and Nilsson, 1994).

Figure 1 shows that the largest relative errors occur primarily in Northern Europe, followed by the Alpine region and Scotland. It could therefore be that the largest *nRMSEs* are related to uncertainties in the simulation of winter precipitation. In climates with a seasonal snow cover, the annual discharge peak usually occurs in spring, when the snowpack that has accumulated over the winter season, melts away. This effect would be exacerbated if a particular river basin contains large lakes and reservoirs – as is for example the case in Finland – that would be able to store a large fraction of the spring runoff, something not simulated by LISFLOOD. In most other areas, the error in the simulated peak discharges seems acceptable.

Contrary to what might be expected, the H12CL run is not giving the lowest error everywhere. The H50CL experiment has the lowest *nRMSE* value at the majority of the stations (91), followed by the H12CL (64) and H25CL (54) runs. A clear geographical pattern cannot be deduced from Fig. 1, although H50CL seems to be slightly more dominant in the western part of the continent. However, in some river basins the differences in *nRMSE* between the three experiments are relatively small, with the H25CL run frequently plotting between the two other runs (Fig. 2). This means that at some gauging stations the model performance is increasing with increasing horizontal resolution of the RCM, while at other stations it is decreasing.

3.2 Comparison between the three control run experiments

The RCM simulations may not reproduce the historical weather but the evolution of the climate system is similar in the three HIRHAM experiments. This is illustrated in Fig. 3 which shows the hydrograph of the Rhine River in an arbitrary year of the control simulations. The discharge peaks as well as the low flow periods are synchronous in the three experiments although the degree of correspondence is not always the same. This feature allows us to calculate more direct statistics to compare the three simulations with each other on a day-to-day basis, such as the Nash-Sutcliffe efficiency coefficient (Nash and Sutcliffe, 1970):

$$E_{N-S} = 1 - \frac{\sum_{t=1}^{n} \left(Q_s(t) - Q_r(t)\right)^2}{\sum_{t=1}^{n} \left(Q_r(t) - \overline{Q_r}\right)^2} \tag{2}$$

where $Q_s(t)$ is the simulated discharge at time t, $Q_r(t)$ a reference discharge (usually the observed streamflow), and $\overline{Q_r}$ the mean reference discharge over the period of interest, with n time steps. The coefficient can range from minus infinity to 1, with higher values indicating better agreement: usually a score of 0.6 or higher is regarded as acceptable. The Nash-Sutcliffe coefficient is often criticised for being heavily influenced by large discharge values (e.g. by Criss and Winston, 2008) but in our case it is an appropriate error measure, as it is sensitive to the differences in the timing and volume of the peak flows. For every grid cell we calculated the E_{N-S} of the H25CL and H50CL model runs, using the H12CL run as reference discharge. In this way the results can be interpreted in terms of how well the two lower resolution experiments approach the high-resolution run. The result is shown in Fig. 4 for rivers with an upstream area of above 1000 km^2.

Figure 4 clearly shows that the correspondence between the discharge simulations is larger in Western Europe than in more continental regions, and larger in major rivers (such as the Danube) than in small tributaries. Furthermore, the agreement with the H12CL simulation is stronger for the 25-km experiment than for the 50-km run. In the H25CL experiment, the Nash-Sutcliffe values are above the acceptable levels in most river basins, except for some smaller tributaries, mostly in the eastern part of the continent. In the H50CL run, on the other hand, the performance of the hydrological model with respect to the H12CL simulation is in general much poorer, except in Western Europe and in very large river basins.

Although it is difficult to generalise, the relatively poor disagreement particularly between the H12CL and H50CL runs in Eastern Europe is primarily due to mismatches in the peak flows. While some discharge peaks are nicely aligned, others may be completely absent in either of the experiments (see Fig. 5 for an example from the Upper Danube at Regensburg, Germany, E_{N-S}=0.15). Tentatively, the larger discrepancies in the more continental areas are due to a larger influence of land-atmosphere interactions on the water cycle (see Seneviratne et al., 2006), which in turn are affected by the higher grid resolution of the 25- and 12-km HIRHAM simulations.

To further explore the relationship between river basin size and the correspondence between the three experiments, we plotted the E_{N-S} values against upstream area in Fig. 6. In small river basins there is obviously a wide range of efficiency values, meaning that in some regions the H25CL and H50CL simulations follow the 12-km

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experiment closely, while in other areas there are considerable mismatches. Above about 30 000 km^2 (~120 000 km^2 in case of the H50CL run) there are, however, very few grid cells where the efficiency is below 0.6, confirming once more that in the larger river basins there is generally a good agreement with the H12CL simulation.

4 Comparison of climate change signal

In the scenario period, the correspondence of the H25A2 and H50A2 experiments with the 12-km simulation is similar to the correspondence in the control period. Although there are differences in detail, the general patterns in the Nash-Sutcliffe efficiency values (not shown) are comparable to those shown in Fig. 4. This means that the agreement with the H12A2 run is stronger in Western Europe than in the east, larger in major rivers than in small upstream catchments, and stronger in the 25-km simulation than in the 50-km run. In fact, the H50A2 experiment seems to be in a slightly better agreement with the high-resolution simulation than in the control period.

But what does this mean for the predicted impact of changed climate conditions on the occurrence of peak flows? To answer this question, we fitted a Gumbel distribution to the 30 Q_{max} values in every river cell in both the control and the scenario period. The Gumbel distribution is a special case of the Generalised Extreme Value (GEV) distribution (Coles, 2001; Katz et al., 2002) with the shape parameter (γ) explicitly set to 0. This approach is the same as that followed by Dankers and Feyen (2008) using the same model set-up. Based on this distribution we then estimated the 100-year return level Q_{100} (i.e. the discharge level that has a 1% chance of being exceeded in a given year) and calculated the relative change in the magnitude of the Q_{100} in the scenario period, relative to its corresponding control simulation. Note that even a small increase in the magnitude of a certain discharge level can already have a substantial impact on its expected return time (Allen and Ingram, 2002).

The result is shown in Fig. 7. The general patterns of the changes in flood hazard are comparable between the three experiments. In Northeastern Europe there is a general decrease in Q_{100}, which is related to a reduction in snow accumulation. In this region the natural flow regime is dominated by a snowmelt runoff peak in spring, but because of the warmer winter temperatures the length of the snow season and the total snow accumulation are considerably reduced in the scenario period. In the rest of the continent we see a more mixed pattern of decreases and (sometimes strong) increases in Q_{100}, reflecting the changes in extreme rainfall amounts rather than in snow accumulation. For a more detailed description of the changes in the H12A2 scenario see Dankers and Feyen (2008).

Although the large-scale response in Q_{100} is comparable between the three experiments, at local scale there can be considerable differences, especially between the 12- and 50-km simulations. It is interesting to note that the H12A2 experiment shows more increases in Q_{100} and less decreases than the H50A2 run (see Table 1). The H25A2 falls between these two but is closer to the 12-km simulation. For example, the H12A2 scenario shows an increase in Q_{100} of more than 5% at 40% of the river grid cells, against 35% in the H25A2 experiment and 25% in the 50-km experiment. Additionally, it can be seen in Fig. 7 that the 12-km experiment not only results in a more widespread increase in Q_{100}, but also in increases that are generally larger in magnitude. If the more frequent occurrence of strong increases in the 100-year return level in the 12-km simulation were realistic, this would mean that using lower-resolution climate simulations leads to an underestimation in the assessment of future flood hazard.

5 Discussion and conclusions

This paper tries to answer the question whether very high resolution climate simulations lead to a better representation of hydrometeorological extremes, and result in a climate signal in flood hazard that is substantially different. With respect to the first issue, when comparing the peak discharges that are simulated by LISFLOOD when driven by three experiments of the HIRHAM model with different horizontal resolution, we found no conclusive evidence for the high-resolution RCM simulation at 12 km to perform generally better than the lower resolution runs at 25 or 50 km (see Fig. 1). However, it would be premature to conclude that the H50CL experiment performs equally well, if not better than the H12CL simulation. Not only does the performance differ from basin to basin, the validation is also complicated by the presence of artificial lakes and reservoirs in most of the rivers that were included in the comparison. Flow regulation is not accounted for in the present LISFLOOD set-up but is likely to affect the timing and magnitude of the observed peak flows. A higher horizontal resolution of the climate model is expected to yield better results because of the better representation of orographic patterns in precipitation and precipitation extremes (see Dankers et al., 2007) but also because of the more explicit spatial patterns in, for example temperature, which should have an impact on the simulation of snowmelt by the hydrological model.

In larger river basins, however, we found that the differences between the three HIRHAM experiments tend to become less important. Above about 30 000 km^2 and 120 000 km^2, respectively, the 25- and 50-km runs generally provide a good approximation of the simulations based on the 12-km climatology. This is because local differences tend to average out at larger scales but also because large floods in major rivers are primarily caused by synoptic-scale low pressure systems rather than small and mesoscale convective systems. As noted by Jacob et al. (2007), a change in resolution of the RCM has a relatively minor impact on the simulation of large-scale climate features.

We also found that the general pattern of change in extreme river discharge in the scenario period, as shown in Fig. 7, is comparable between the three experiments. Yet, at the scale of individual river basins there may be important differences. If we assume that the 12-km HIRHAM simulation provides a more realistic representation of changes in extreme rainfall, the use of lower-resolution climate simulations may lead to an underestimation of future flood hazard, and in places even to a climate signal of opposite sign. This implies that results obtained with lower-resolution RCM simulations should be interpreted with care, as the grid scale of the climate model adds to the

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uncertainty in the simulations.

The grid scale of climate simulations is, however, only one source of uncertainty and high-resolution runs are computationally more expensive. Other sources of uncertainty include the scenario of future greenhouse gas emissions, the regional climate model formulation, as well as the global model that is used to drive the RCM (Dankers and Feyen, 2009). Furthermore, climate simulations typically show a large, decadal-scale variability that may obscure the climate change signal produced by a change in external forcing, i.e. a rise in greenhouse gas concentrations. Using an ensemble of RCM simulations from the PRUDENCE project, Dankers and Feyen (2009) found that, due to this internal variability, differences in flood levels may arise when comparing two 30-year periods with each other, even without a change in climate. This calls for more probabilistic climate scenarios consisting of multiple realisations of the current and future climate state, and these are obviously more easily obtainable with lower-resolution models.

The answer to the question on the added value of very high RCM experiments in impact studies of hydrological extremes therefore depends on the objective and scale of the study. In major river basins the performance of medium-resolution RCM simulations with a grid scale of 25 or 50 km seems to be sufficient and allows for multiple realisations in order to diminish the influence of natural variability on the simulated climate response. At smaller scales or in areas with large topographical differences, a high resolution RCM run may provide a more realistic simulation of precipitation extremes and therefore less bias in the simulated flood hazard. In either case, one has to be aware of the uncertainties in the simulated climate impact on discharge extremes arising from different sources, and a thorough validation of the simulations is therefore essential.

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Table 1. Comparison of changes in Q_{100} in the scenario period in the three experiments. The numbers are percentages of grid cells with an upstream area of more than 1000 km^2.

		H12A2			
		>5% decrease	little change	>5% increase	*Total*
H25A2	>5% decrease	34	5	6	*46*
	little change	8	6	6	*20*
	>5% increase	3	5	26	*35*
H50A2	>5% decrease	36	8	12	*56*
	little change	4	5	11	*20*
	>5% increase	3	4	18	*25*
Total		*44*	*16*	*40*	

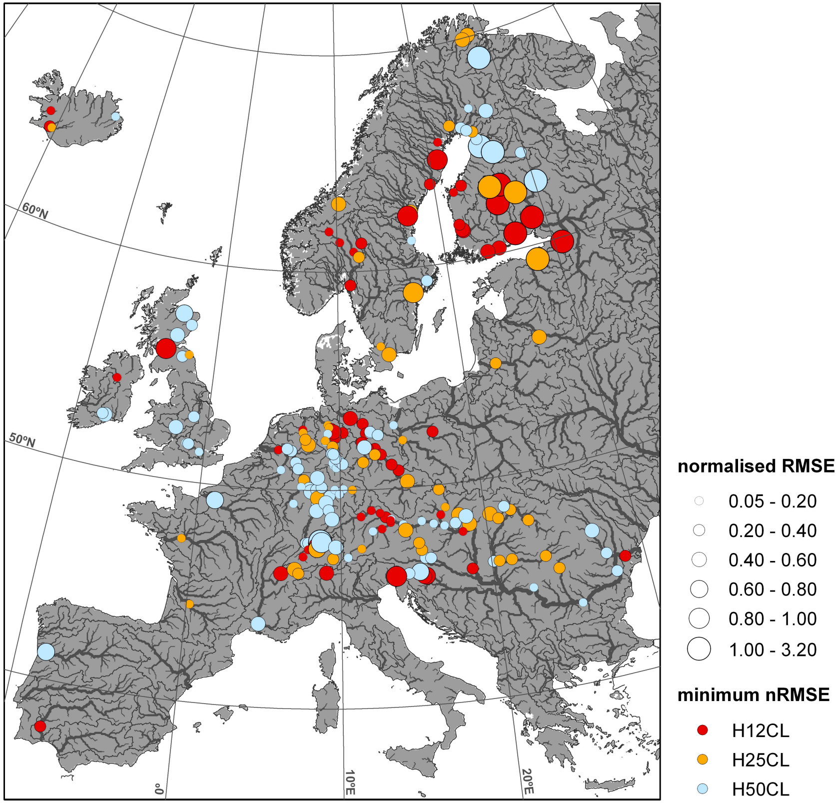

Fig. 1. Normalised root mean square error (*nRMSE*) of the observed and simulated annual maximum discharges (Q_{max}), ranked from high to low, at 209 discharge gauging stations in Europe. The size of the symbols relates to the magnitude of the *nRMSE*, the colour indicates which of the three model experiments has the lowest error.

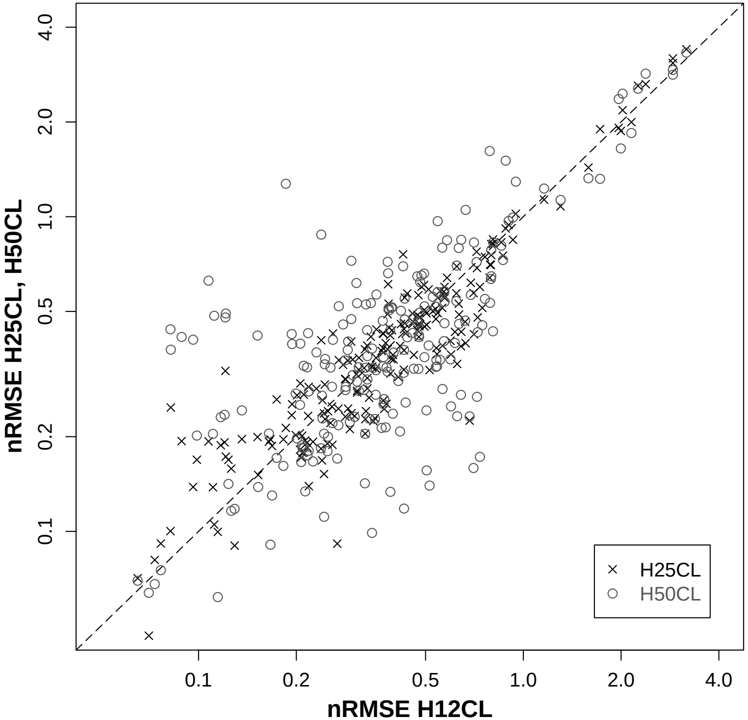

Fig. 2. Scatterplot of the *nRMSE* of the H25CL and H50CL model runs against the *nRMSE* of the H12CL run, at the 209 gauging stations used for validation.

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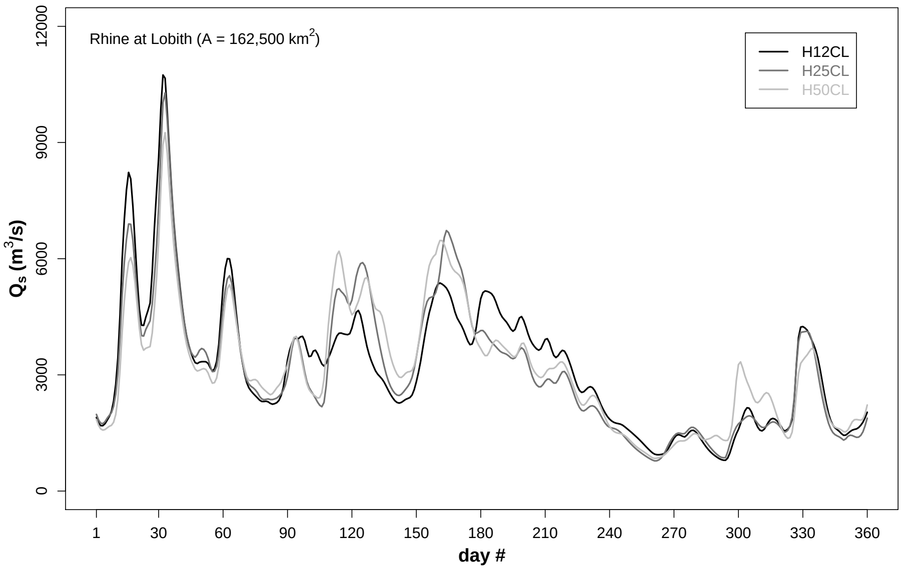

Fig. 3. Simulated discharge in the three model experiments of the Rhine River at Lobith (The Netherlands) in an arbitrary year (y=10).

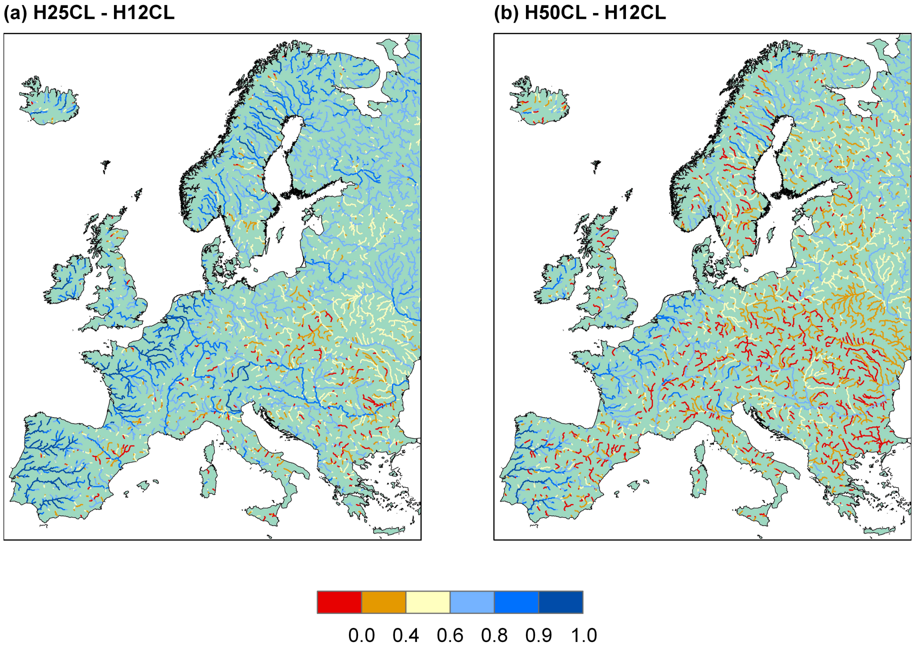

Fig. 4. Nash-Sutcliffe efficiency coefficient of the H25CL **(a)** and H50CL **(b)** model runs, using the H12CL run as reference. Results are shown for grid cells with an upstream area of more than 1000 km^2.

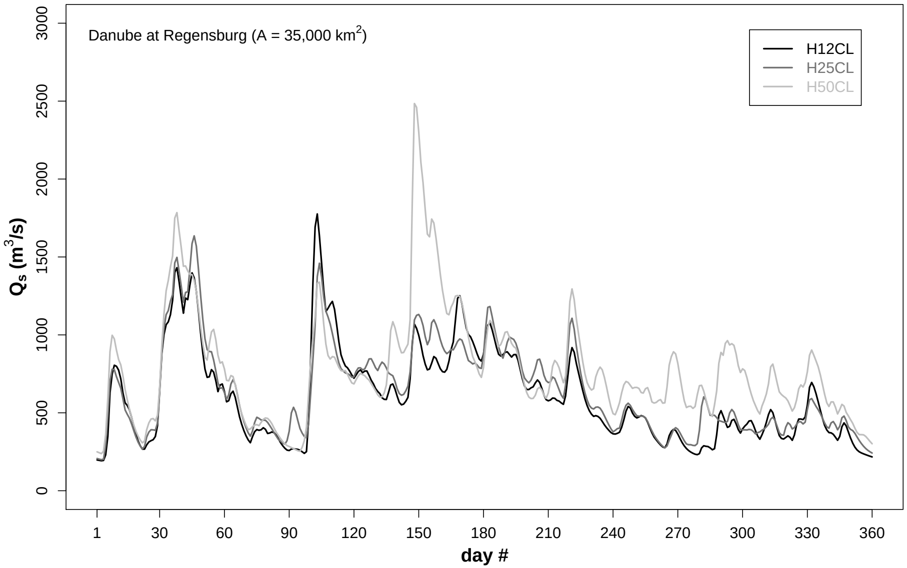

Fig. 5. Simulated discharge in the three experiments of the Danube at Regensburg (Germany) in an arbitrary year (y=6).

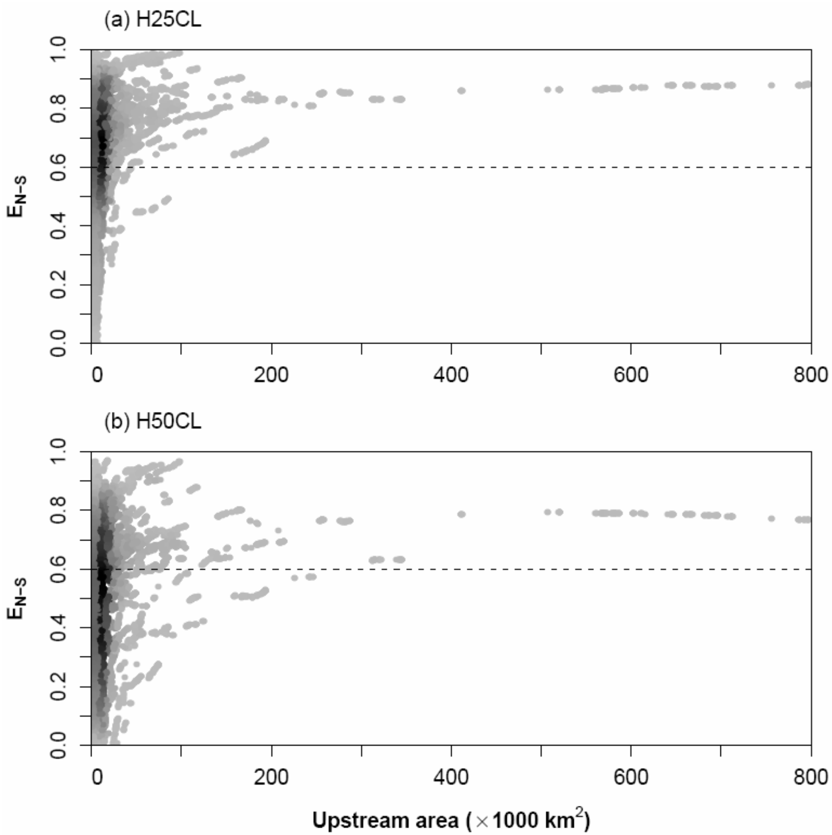

Fig. 6. Values of the Nash-Sutcliffe coefficient (E_{N-S}) of the H25CL **(a)** and H50CL **(b)** experiments, using the H12CL run as reference, plotted against upstream area. Darker shades represent a larger clustering of points.

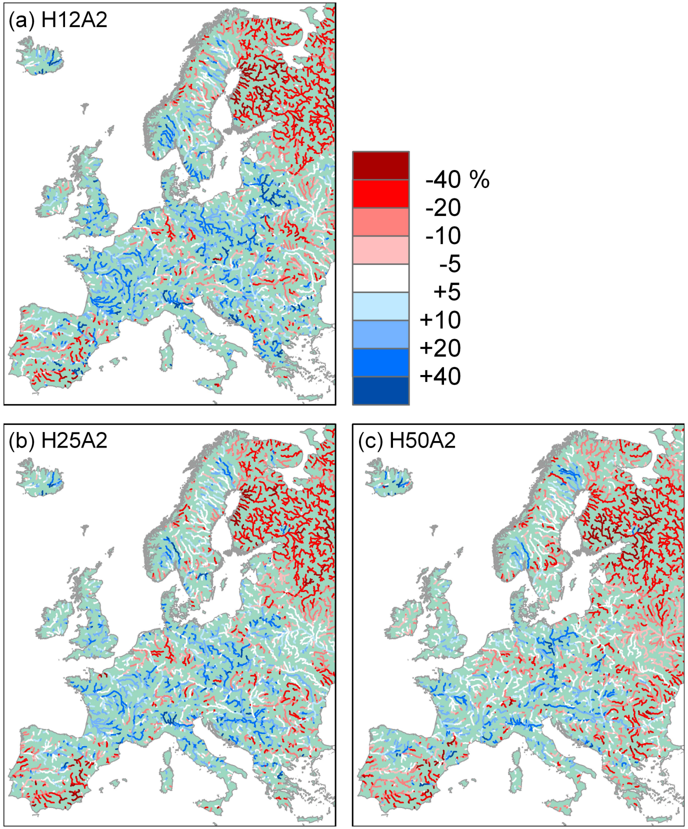

Fig. 7. Change in the magnitude of the 100-year return level (Q_{100}) in the scenario period (2071–2100) compared to the control period (1961–1990), in the H12A2 **(a)**, H25A2 **(b)** and H50A2 **(c)** scenarios. Estimates of Q_{100} are based on a Gumbel distribution.

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