6 Changes in Intense Precipitation in Europe

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6.1 INTRODUCTION

Intense precipitation may have disastrous impacts on economic and social conditions, forcing river and flash floods, and causing strong local anomalies of soil moisture and groundwater storage. This is especially important for densely populated Europe, which is strongly exposed to hydro-climatic extremes, to the extent that the existing risk and water management practices have major difficulties in coping with the ongoing changes. From a climate change perspective, extreme precipitation is believed to be one of the key mechanisms re-organizing the global water cycle under anthropogenic warming conditions (Allan & Soden, 2008).

Precipitation represents one of the most uncertain characteristics in climate models due to the poor quality of parameterizations used in present model configurations. However, there are also many uncertainties inherent in climate analyses of observed precipitation (both *in situ* and remotely sensed). The reasons for these uncertainties reside primarily in (but are not limited to) the high variability of precipitation at all spatial and temporal scales. Being strongly clustered in space and in time, precipitation is an event-like phenomenon with no simple parallel in the other climate variables, such as temperature and pressure. Thus, the standard methodologies used in the analysis of classical meteorological scalars may not necessarily be effective for the analysis of precipitation. The uncertainties drastically increase when we deal with extreme precipitation. In this case, besides the general difficulties when analysing precipitation, we face problems of definition, because it is often uncertain what can be termed as extreme precipitation in different areas and for different tasks. For example, Fig. 1 shows daily precipitation time series for two stations in Scandinavia in summer 1982. If we consider the threshold of 10 mm/day (widely used for definition of heavy precipitation events) and analyse precipitation values above this threshold for the Swedish station Stenslese, we will obtain several events which can be classified as extreme. At the same time, for the Norwegian station Bulken, rainfalls exceeding in magnitude the same threshold (10 mm/day) will be classified locally as ordinary precipitation events. This shows that, even for regional studies, one should be very careful when quantifying extreme precipitation events.

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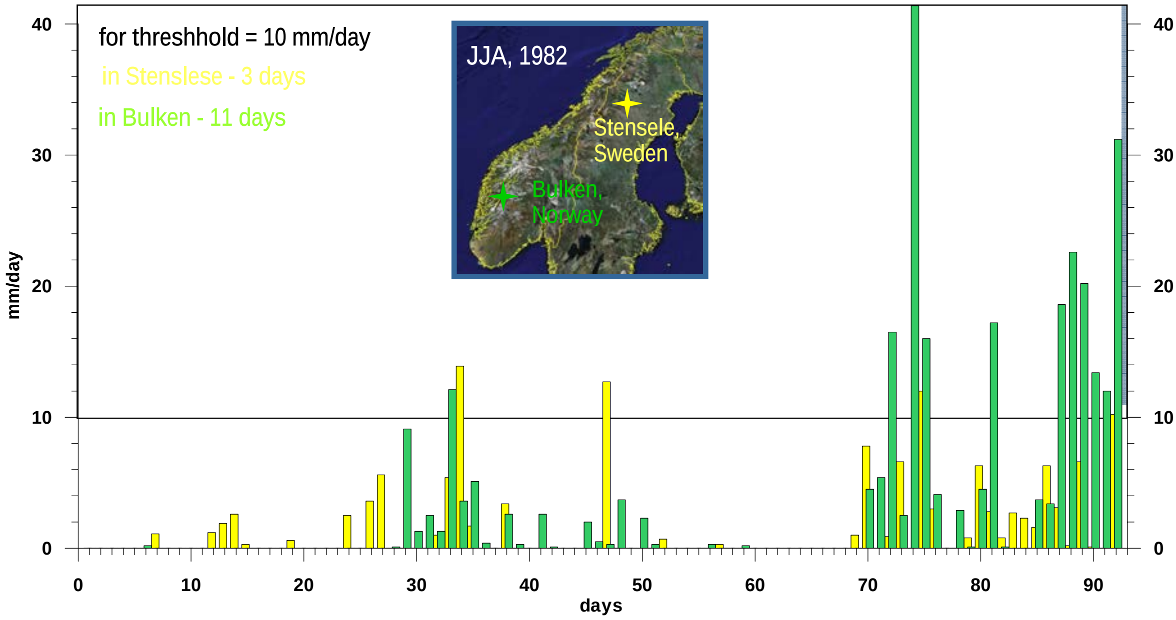

Fig. 1 *Daily precipitation for stations Bulken in Norway (green) and Stensele in Sweden (yellow) for summer 1982.*

6.2 DATA FOR ESTIMATION OF INTENSE PRECIPITATION

Most of the discrepancies mentioned in the Introduction (6.1) are to a large extent caused by uncertainties on the observational side and in the analysis techniques employed. Considering the analysis of precipitation variability, we can mention many data sources that are particularly valuable for such an analysis. If we do not discriminate between the data sets with respect to their temporal coverage and quality, we can state that, for instance, mesoscale model data, ground-based radars and raingauges can properly resolve the scales required. However, if we require the coverage of at least several decades for studying climate variability and change, we will remain with global re-analyses on the modelling side and with only the data from daily raingauges on the observational side.

Re-analysis precipitation products (e.g. Kalnay *et al*., 1996; Uppala *et al*., 2000; Kistler *et al*., 2001; Kanamitsu *et al*., 2002) provide precipitation estimates along with basic meteorological variables over the globe with high temporal resolution. However, precipitation appears to be one of the most uncertain parameters in the re-analyses due to the still-poor skill of operational Numerical Weather Prediction (NWP) models in accounting for all the important physical mechanisms that affect the atmospheric water cycle and precipitation generation. Thus, these data must be extensively validated against alternative data sources (first of all raingauges) and compared to each other (e.g. Zolina *et al.*, 2004) before they can be reliably used for the analysis of precipitation variability at the climate scale.

Given the above mentioned reasons, station raingauge data should be considered as the most reliable source of information for estimation of intense precipitation over Europe. Although continental-scale “climate quality” raingauge records are believed to be properly validated and standardized, their resolution is too coarse to allow for a sufficiently detailed and consistent analysis of the variability in extreme precipitation and its mechanisms due to the multi-scale characteristics of rainfall in general. Furthermore, station records are influenced by sampling inhomogeneities affecting the estimates of extreme precipitation.

Figure 2 shows the spatial distribution of raingauge stations with data available from different collections over Europe. The most widely-used collection was developed by merging three data sets: the Royal Netherlands Meteorological Institute (KNMI) collection, known as the European Climate Assessment (ECA) daily data set (Klein Tank *et al*., 2002; Klok & Klein Tank, 2009). After the recent updates (Klok & Klein Tank, 2009) this data set consists of 1558 synoptic stations, which cover periods of one or two decades up to more than 100 years. In Zolina *et al.* (2009), we corrected some biases in 57 ECA stations (over European Russia, Belorussia and Ukraine for the 1990s and 2000s) due to incorrect record decoding, and 32 new stations were added. An additional source of information is represented by SYNOP stations assembled in the merged NCDC and Russian Metoffice collection (Fig. 2). However, the joint use of ECA data and SYNOP stations still remains problematic due to the different quality control procedures applied in these data sets.

There are also national data collections maintained by national meteorological offices and covering individual states with very high spatial resolution. For instance, the operational raingauge network of the German Weather Service (DWD) (Fig. 2(b)) covers Germany with a very high spatial resolution, and is one of the densest and most properly maintained regional precipitation networks (Zolina *et al.*, 2008). It consists of 11 617 stations of which 5454 have been digitized, controlled and included in a digital database (MIRAKEL-Datenbank), while the other 6163 are still in the process of continuous pre-processing at DWD and its partners. All DWD precipitation stations are equipped with Hellmann raingauges with a collecting surface of 200 cm^2. All rain-gauges are shielded with a special cover to protect against evaporation loss, and are

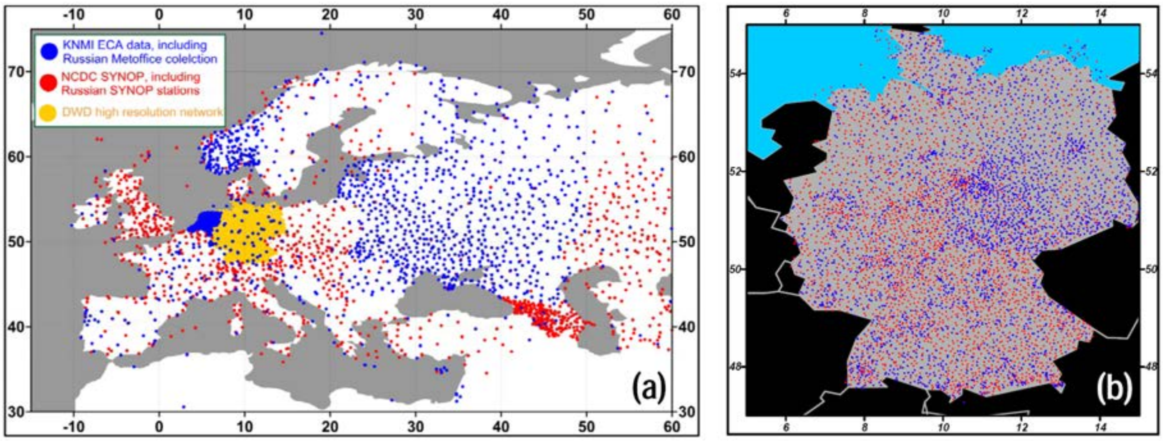

Fig. 2 *(a) European precipitation observation database combined from different collections. (b) Operational raingauge network of DWD. Red points show the rain-gauges covering the period 1950–2004. The blue dots show the raingauges which do not cover 1950–2004 (adapted from Zolina et al., 2008).*

equipped for snow with built-in heating devices, controlled by thermostats, including power supply units. These devices are typically activated at the beginning of the cold season, when each instrument is additionally supplied with a snow cross and a second lid. These prevent possible blowing of the snow. For snow precipitation, the water level of the melted precipitation is calculated. The raingauges are installed and exposed such that their collecting surface is 1 m above ground. The accuracy of the precipitation measurements is 0.1 mm. In the late 1990s, the change from the manual reading procedures at DWD raingauges to automatic reading began by installation of pluviometer devices. These provide estimates of daily (and much higher resolution) precipitation based on weighing. Implementation of this transformation of the reading methodology did not introduce any change of the times to which measurements are attributed.

Unevenly distributed missing values in daily precipitation records may cause artificial time-dependent biases and, therefore, affect estimates of inter-annual variability on both decadal and centennial time scales. In order to quantify these effects on estimated extreme precipitation, we performed a sensitivity study based on a Monte Carlo simulation of missing values following different sampling models (Zolina *et al*., 2005). We selected about 30 stations with gap-free daily records during the time period 1900–2002. For these time series, we simulated undersampling according to the gap structure derived from an analysis of all European stations. Then we performed an analysis of the sensitivity of extreme precipitation indices to the number of missing values. For all indices the impact of sampling on linear trends is more pronounced for mountain, island and coastal stations, which are more strongly affected by convective precipitation. On the time scale of interannual variability, gaps may locally produce 10 to 50% of the interannual variability that is not explained by the gap-free time series of extreme precipitation. A general conclusion of Zolina *et al.* (2005) was that, for the unbiased analysis of long-term variability in extreme precipitation, it is necessary that seasonal time series have no more than 30% of missing daily values.

6.3 ABSOLUTE PRECIPITATION EXTREMES OVER EUROPE

As mentioned above, estimation of extreme precipitation requires more metrics compared to the other meteorological variables. If we consider absolute values, the use of the raw data is the simplest and not necessarily the worst method. We can calculate maximum precipitation, peak-over-threshold, or simply precipitation intensity. However, for a robust statistical estimation of extreme precipitation, a general practice is to apply theoretical parametric probability density distributions. In this paradigm there is a conceptual problem as to whether it is more appropriate to use initial value distributions (IVD, like the Gamma distribution) built for the whole sample, or extreme value distributions (EVD) for the extreme precipitation events only (e.g. Friederichs & Hense, 2007; Friederichs, 2010; Maraun *et al.*, 2010).

According to Maraun *et al.* (2010), distributions built for the whole sample (IVD) may not necessarily capture the extremes accurately enough. On the other hand, results based on EVD, for example the Generalised Extreme Value distribution (GEV), strongly depend on the choice of threshold and may significantly overestimate extremes. Although this does not contradict in general the extreme value theorem (see e.g. Kharin & Zwiers: 2000; Friederichs, 2010), it does raise a question about the impact of interdependency of the adjacent peak-over-threshold estimates derived from daily time series. Even if the block maxima (asymptotically following a GEV distribution, Friederichs, 2010) are considered, the adjacent values can also be interdependent.

Figure 3(a) and (b) shows linear trends over the last 110 years and over the last 60 years in 95th percentile of daily precipitation over Europe for winter and summer seasons. During the 20th century there has been an upward tendency in the magnitudes of extreme precipitation over Europe, with the stronger trends in winter (Fig. 3(a); 4 to 10% per decade) compared to summer (Fig. 3(b); typically less than 6% per decade). This implies that during a 100-year period, heavy precipitation corresponding to the 95th percentile increased from 1 to 3 mm/day per decade in Central Europe. Interestingly, the strongest centennial trends are identified for eastern European Russia, where trend magnitudes are 20–30% higher than in Western Europe.

During the last 60 years (Fig. 3(c) and (d)), extreme winter precipitation intensified by about 2–4% per decade in Central Europe and by up to 6–8% per decade in Western Europe (Fig. 3(c)). For the summer season (Fig. 3(d)) changes in the intensity of extreme precipitation were still evidently positive in Eastern Europe, but

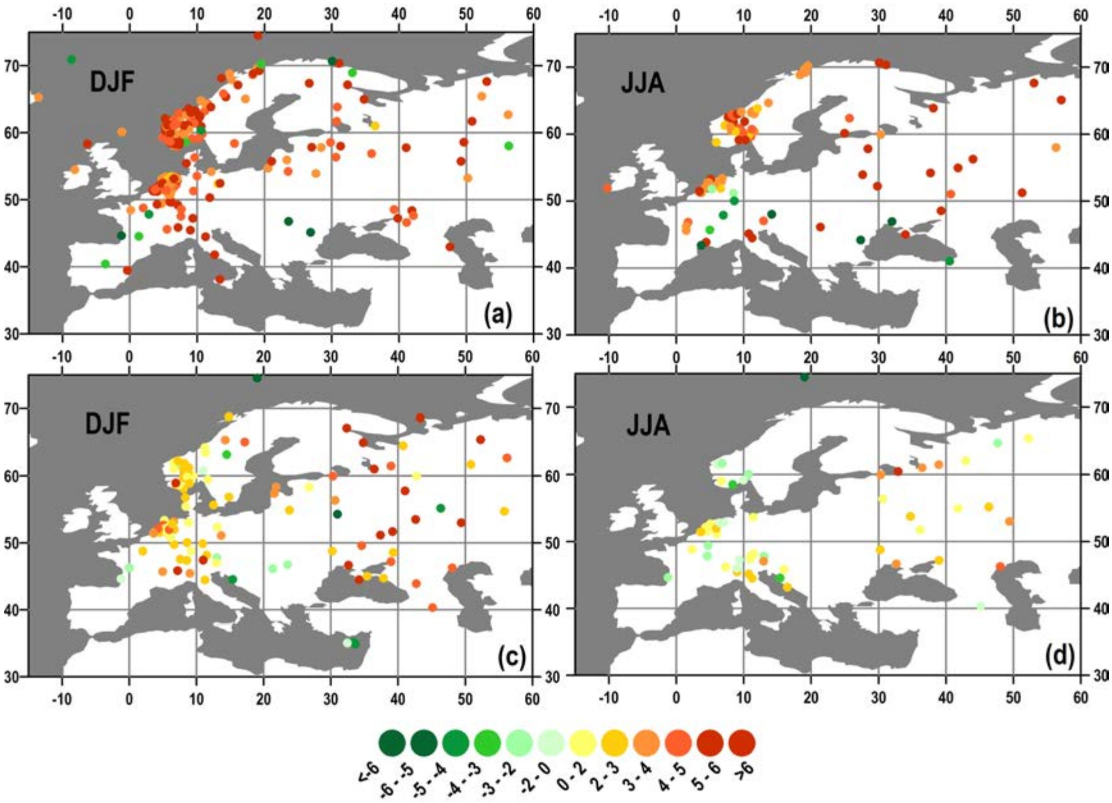

Fig. 3 *Linear trends over the last 110 years (1901–2010) (a,b) and over the last 60 years (1951–2010) (c,d) in 95th percentile of daily precipitation over Europe for winter (a,c) and summer (b,d) seasons. Only trends that are significant at the 95% level have been shown.*

demonstrated negative trends of about 1–3% per decade in many locations of Central Western Europe.

Upward tendencies for the last century and the last five decades were confirmed in regional studies by Frei & Schär (2001), Groisman *et al.* (2005), Brunetti *et al.* (2004, 2006), Moberg *et al.* (2006), Alexander *et al.* (2006) and others, who also contributed to the IPCC Fourth Assessment Report (AR4) (Trenberth *et al.*, 2007), which confirmed an increasing number of heavy precipitation events over the last 50 years over Europe.

A very remarkable feature of European precipitation for many regions is the quantitatively (and, locally, qualitatively) significantly different linear trends in the mean intensities and in the magnitudes of precipitation extremes. Figure 4 shows 55-year long variability in the mean precipitation intensity and in the 95th percentile of precipitation in Zagreb. Interestingly, mean intensities demonstrated a clear downward tendency, while the frequency of extreme precipitation has grown by about 3% per decade; both trends are statistically significant.

Importantly, however, the estimates of extreme precipitation trends over Europe are spatially quite inhomogeneous. This is shown in Fig. 3 and was also shown by many published spatial patterns of linear trends for the last 50 years. Frei & Schär (2001) noticed a strong spatial inhomogeneity in the trend estimates, even for the Swiss Alps, on the centennial time scale. Likewise, the gridded extreme precipitation indices from Alexander *et al.* (2006) do not reveal a robust pattern of tendencies in extreme precipitation over Europe. Spatial trend inhomogeneity is especially pronounced in mountain regions (Beniston, 2005). Similar noisy trend patterns were reported for the North American continent (Kunkel *et al.* 1999, 2003; Groisman *et al.*, 2005).

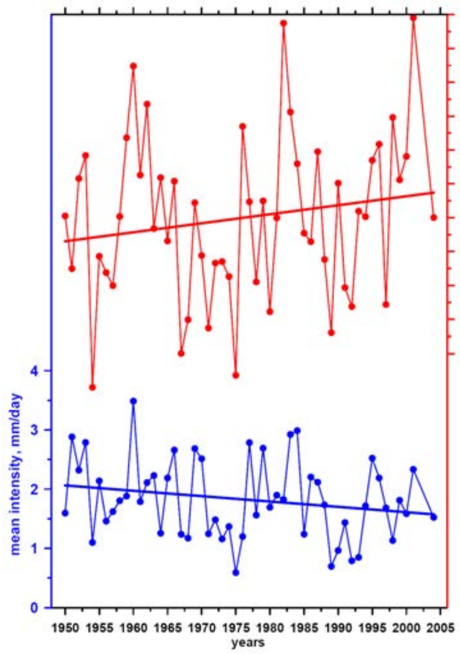

Fig. 4 *Time series in mean precipitation intensity (blue) and in 95th percentile of precipitation (red) in Zagreb for the period from 1950 to 2005.*

Another remarkable feature of long-term interannual variability of extreme precipitation is the seasonality of long-term trends in extreme precipitation. Seasonality in trends of mean European precipitation and temperature were found by Shabalova & Weber (1999), Datsenko *et al.* (2001) and Zveryaev (2004). Generally, these should imply seasonality also of trends for extremes. Recently, Moberg *et al.* (2006) analysed winter and summer trends in mean and extreme precipitation indices over Europe during 1901–2000. They found significant increases in heavy and extreme winter precipitation, while linear trends in summer were found to be mostly insignificant and regionally dependent. However, to find the evidence of seasonality, it is necessary to use dense observational networks. Seasonal maps (Fig. 3) show some signatures of trends of different signs in European precipitation in central Western Europe; however, they can hardly be considered as a robust pattern.

For quantification of seasonality signals we analysed more than 5000 raingauges over Germany (Zolina *et al.*, 2008). Only 2125 of the total of 5454 stations were selected for the analysis when applying the data completeness requirements (Zolina *et al.*, 2005). From this subset we further excluded 576 stations (primarily from the former German Democratic Republic) which lack information about the observation techniques applied. The final selection has 2125 stations that have considerably fewer gaps compared to the complete data set for the period 1950–2004.

Our analysis indicated a clear seasonality of extreme precipitation over Germany (Fig. 5). The winter season trend pattern (Fig. 5(a)) shows positive trends everywhere over northern and western Germany. More than 1800 locations over Germany indicate positive trends, while about 90 stations show downward tendencies. Positive trends at the 95% significance level were obtained for 666 of the 2125 raingauges considered. Only 17 locations indicate significantly negative linear trends for the 95th percentile of daily precipitation (p_{95}); the latter are mainly located in southern and southeastern Germany. Increases in p_{95} above 11% per decade are observed in central western

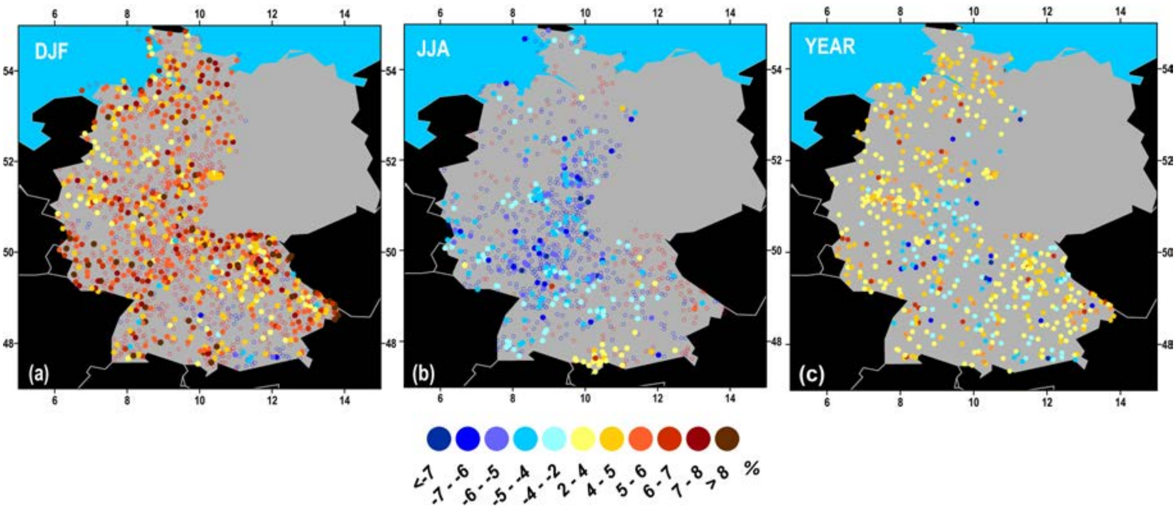

Fig. 5 *Linear trends (% per decade) of the 95th-percentile of daily precipitation for: (a) winter, (b) summer and (c) annual time series, for the period 1950–2004. Small circles show all trend estimates and large circles denote the locations where trends are significant at the 95% level.*

Germany and in the southeastern mountain regions, which result in an increase of heavy precipitation of about 1.5–2.5 mm/day and 3–4 mm/day per decade, respectively. Summer trends in p_{95} (Fig. 5(b)) show opposite signs compared to those found for winter (as well as in spring and autumn) with decreases of as much as 7% per decade (2–4 mm/day per decade) in the central part of western Germany. About 1800 stations in total have negative linear trends and only 300 exhibit positive linear trends. Statistically significant positive summer trends form a local cluster in southern Germany with a 2 to 4% increase per decade.

To estimate trend significance, in addition to standard tests (e.g. Student *t*-test, Wilcoxon test), we applied non-parametric tests (Mann-Kendall test) and estimated the confidence limits according to Hayashi (1982). Furthermore, we assessed the potential impact of the variable beginning and ending years of the records on the trend estimates and found that our results are insensitive to this variability. Finally, we estimated the field significance of the trend patterns following Livezey & Chen (1983) from the binominal distribution and according to the Walker test (Wilks, 2006). The trend patterns keep the field significance and substantiate the robustness of positive trends for the winter, spring and autumn seasons and of the negative trends for the summer period.

The spatial distributions of linear trend estimates from the annual time series (Fig. 5(c)) are rather noisy, with neighbouring patches having opposite trends. Obviously extreme precipitation characteristics estimated from the annual time series do not show a clear signal of either sign over the whole domain. This is especially true for central and southwestern Germany where stations indicating trends of the opposite signs are mixed on spatial scales of tens to a hundred kilometres.

Our analysis of seasonality of extreme precipitation over Germany also allowed us to answer the question as to whether the observed seasonality in extreme and heavy precipitation characteristics is also related to changes in weak and moderate precipitation. To this end, we estimated trends for the 10% quantiles of the fitted gamma PDF of daily precipitation for selected regions. Quantifying the role of precipitation with different intensities (weak, moderate and strong precipitation) allows the identification of the disproportional (Easterling *et al.*, 2000; Groismann *et al*., 2005) changes in precipitation extremes. With some minor methodological modifications, this approach was used by Gershunov (1998) and Karl & Knight (1998) for North American raingauges, by Brunetti *et al.* (2004, 2006) for the Italian collection of daily station measurements, and by Maraun *et al.* (2009) for British stations. In Fig. 6(a) and (b) we show the distribution of linear trends for different classes of precipitation intensity averaged over western Germany. The precipitation intensity of 70% and higher percentiles is characterized by a statistically significant 2–3% decadal increase. This tendency is associated with a pronounced decrease of the intensity of weak precipitation (<20% percentiles) of more than 2% per decade. Thus, increasing extreme precipitation in winter is associated with a modification of its PDF. During the summer season (Fig. 6(b)), statistically significant negative trends are observed for all precipitation classes, ranging from 1.5% per decade for heavy precipitation to about 3.5% per decade for weak precipitation.

To demonstrate that clear evidence of trend seasonality can be only effectively captured by dense observational networks, we show, in Fig. 7, results of the analysis of trends at 25 stations of the ECA collection covering our area. The trend signs in p_{95} are primarily positive in winter and primarily negative in summer. However, in winter,

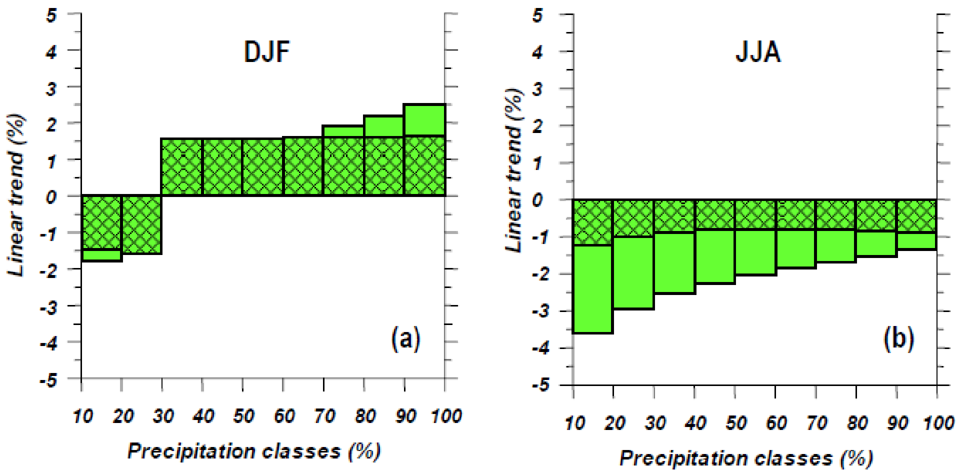

Fig. 6 *Linear trends (% per decade) of precipitation intensity for different precipitation classes averaged over all stations for: (a) winter and (b) summer.*

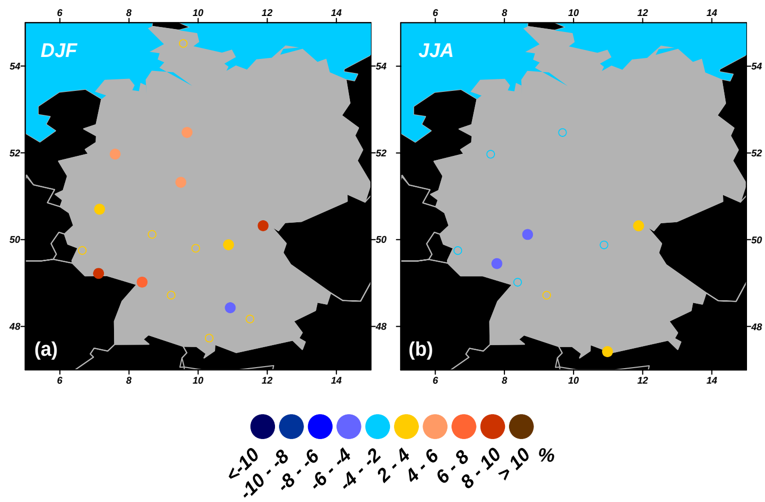

Fig. 7 *Linear trends (% per decade) in the 95th percentile of daily precipitation for: (a) winter and (b) summer, derived from the ECA collection of raingauges for the period 1950–2004. Trends significant at 90% level are shown by closed circles.*

positive trends at the 90% level of 4 to 10% per decade are observed for eight locations. This does not guarantee the field significance even at the 90% significance level according to both the binary and the Walker test. Remarkably, in summer, significantly negative (at the 90% level) trends in heavy precipitation were identified in only two of 25 locations. These estimates are in agreement with Klein Tank & Können (2003), who noted quite noisy patterns of linear trend estimates using the ECA collection. Thus, the high quality ECA collection cannot be used for testing seasonality in temporal changes of precipitation characteristics due to its coarse resolution.

6.4 CLIMATE TENDENCIES IN THE RELATIVE PRECIPITATION EXTREMENESS OVER EUROPE

When analysing heavy precipitation, we also need to distinguish between absolute extremes and “extremeness”, or the contribution of the most wet days to the total precipitation. The estimation of extremeness allows us to account for the fractional contribution of the wettest days to the precipitation totals, and helps us to distinguish between changing mean precipitation and the changing occurrence of the wettest days in forming precipitation extremes. For instance, monthly or seasonal precipitation total may remain stable while heavy precipitation events can occur on a varying sequential number of wet days, implying changing relative extremeness of precipitation. Approaches to quantify “extremeness” have so far been based on empirical extreme precipitation indices (Klein Tank & Können, 2003; Moberg *et al.*, 2006; Alexander *et al.*, 2006). These indices quantify the precipitation contribution to the total, which is caused by days with precipitation exceeding a given percentile (e.g. 75% or 95%). Such indices are, e.g. *R75tot* and *R95tot*. According to Klein Tank & Können (2003), *R95tot* is computed as:

$$R95tot = \frac{\sum_{n=1}^{N} R_n \middle| R_n > R95}{\sum_{n=1}^{N} R_n} \qquad (1)$$

where R_n is the daily precipitation on wet day n ($R \geq 1$ mm), N is the number of wet days, $R95$ is the 95th percentile of daily precipitation estimated for the whole analysing period. Note here that the choice of the reference period can seriously affect the estimates of relative extremeness. In Klein Tank & Können (2003), the 1961–1990 period was chosen as a reference for most IPCC TAR and AR4 assessments, which were based on the data records typically covering the period from the 1950s to 2000s. However, for the analysis of data sets with different intervals (e.g. the whole 20th century, the last 20 years), this period may not necessarily be representative, leading to differences between the estimated indices. To avoid this, the choice of a reference period should be always coordinated with the actual period of observations available. Furthermore, the estimation of indices for relative extremeness for seasonal or even monthly time series is severely hampered by the finite number of wet days per season or month (Zolina *et al.*, 2009). Figure 8 shows time series of the *R95tot* index computed for the station Sodankyla (Finland). Due to the impact of the finite number of wet days used for computation, estimates of *R95tot* strongly fluctuate and may even fall to zero (22 of 50 winters during the 50-year period). The uncertainty associated with the finite number of wet days in raw time series may seriously affect the estimation accuracy of *R95tot*.

Figure 8(a) shows the number of cases when *R95tot* computed using annual, seasonal and monthly time series falls to zero for different percentiles for all European stations. Even for the annual time series this index cannot be accurately estimated for percentiles higher than the 95th percentile. When computed for seasons, *R95tot* frequently falls to zero before the 85th percentile.

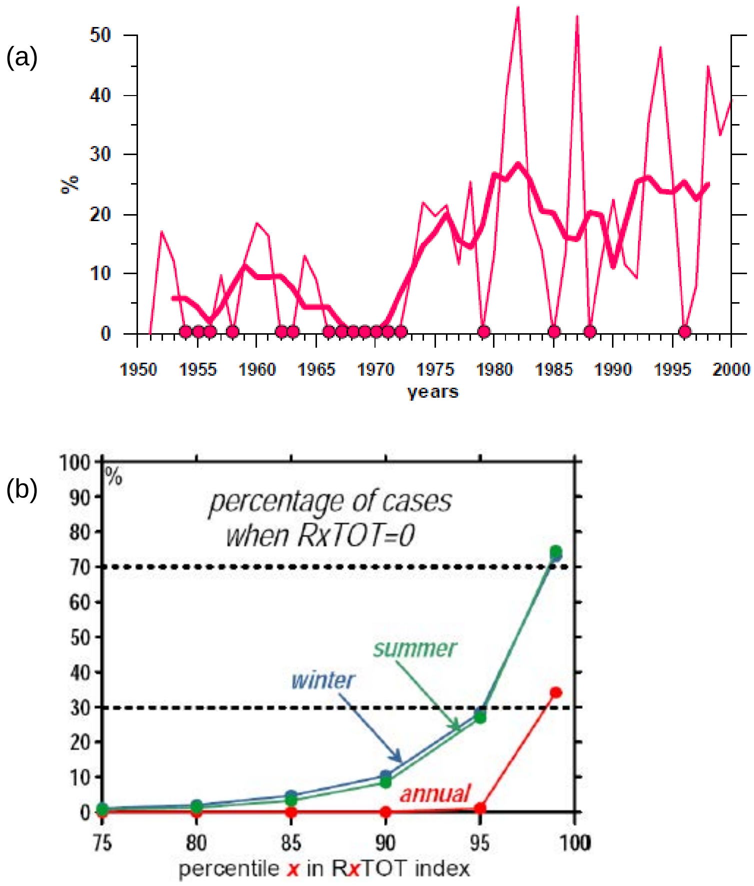

Fig. 8 *(a) Time series of R95tot for station Sodankyla (Finland) during winter seasons from 1951 to 2000. The thick line shows the 5-year running mean values, circles indicate years when the index becomes 0. (b) The number of cases when Rxtot falls to zero for different percentiles x and different seasons.*

The existing approaches to estimating relative extremeness of precipitation have limitations due to the finite number of wet days (Klein Tank & Können, 2003). As a result, the trend estimates based on the traditional index of the relative extremeness, such as *R95tot*, widely used in IPCC Fourth Assessment Reports (Trenberth *et al.*, 2007), is plagued by large uncertainties. To avoid these limitations and to establish a universal index of relative extremeness, a probability distribution of the fractional contribution of daily precipitation amounts to the total was derived by Zolina *et al.* (2009). This distribution, called DFC (Distribution of Fractional Contribution) was further applied to study European precipitation extremes. In Zolina *et al.* (2009) we derived the distribution of the partial contribution of daily precipitation to the monthly/seasonal total, based on the gamma PDF. The gamma distribution is a positively-skewed, two-parameter distribution bounded by zero on the left; its PDF $F(x)$ is given by:

$$F(x) = \frac{\left(\frac{x}{\beta}\right)^{\alpha-1} e^{-\frac{x}{\beta}}}{\beta \Gamma(\alpha)}, \quad \alpha > 0, \quad \beta > 0 \tag{2}$$

where $\Gamma(\alpha)$ is the gamma function, α is the non-dimensional shape parameter determining the skewness of the PDF, and β is the scale parameter, which holds the dimension of the variable analysed and steers the stretch and squeeze of the PDF. The mean intensity of precipitation equals $\alpha\beta$, implying that shape and scale parameters are dependent.

Despite several studies where distributions different from gamma were used for the approximation of PDFs of daily precipitation (e.g. Weibull distribution), the gamma PDF is widely accepted as the most appropriate approximation (Wilks, 1995). A detailed analysis of the applicability of the gamma PDF to European precipitation including parameter estimation is given, for example, in Semenov & Bengtsson (2002) and Zolina *et al.* (2004). We note that for further consideration we will require the distribution for the whole sample of precipitation (IVD) and not only the distribution of extreme values. Given this concept, the gamma distribution is unconditionally most appropriate. From the gamma PDF (equation (2)), we can derive a distribution for the ratio between daily precipitation and precipitation total during a given period, e.g. a month or a season:

$$P(y) = P\left[(x_i \Big/ \sum_{i=1}^{n} x_i) < y \right] \tag{3}$$

where x_i, $i = 1, \ldots, n$ is daily precipitation, n is the number of wet days (the size of the sample), and $y_i = x_i \Big/ \sum_{i=1}^{n} x_i$. The argument y in equation (3) is bounded by [0, 1] and depends on α and n, but is independent of β. Mathematical transformations (Zolina *et al.*, 2009) lead to the following explicit form for $P(y)$:

$$C(y) = P(Y < y) = \frac{\Gamma(n\alpha)}{\alpha \Gamma[(n-1)\alpha] \Gamma(\alpha)} y^{\alpha} (1-y)^{(n-1)\alpha} F_2^1(1, n\alpha, \alpha+1, y) \tag{4}$$

where $F_2^1(a, b, c, y)$ is the Gaussian hypergeometric function. The PDF of this distribution has the following form:

$$F(y) = \frac{\partial C(y)}{\partial y} = \frac{\Gamma(n\alpha)}{\Gamma[(n-1)\alpha] \Gamma(\alpha)} y^{\alpha-1} (1-y)^{(n-1)\alpha-1} \tag{5}$$

Using equations (4) and (5), we can compute an analogue of the *R95tot* index, derived from the theoretical distribution (*R95tt* herein) and compare it with the *R95tot*.

According to Fig. 9, the *R95tot* estimates are, on average, higher by about 8% of the mean values compared to *R95tt*, when derived form the annual time series. The seasonal *R95tot* estimates are somewhat lower than *R95tt*, especially in winter. The largest winter differences of 4–5% (25% of mean values) are observed in central Western Europe. During summer, the differences between *R95tt* and *R95tot* are smaller (compared to winter) with a few locations in Eastern Europe where *R95tot* is even slightly higher than *R95tt*.

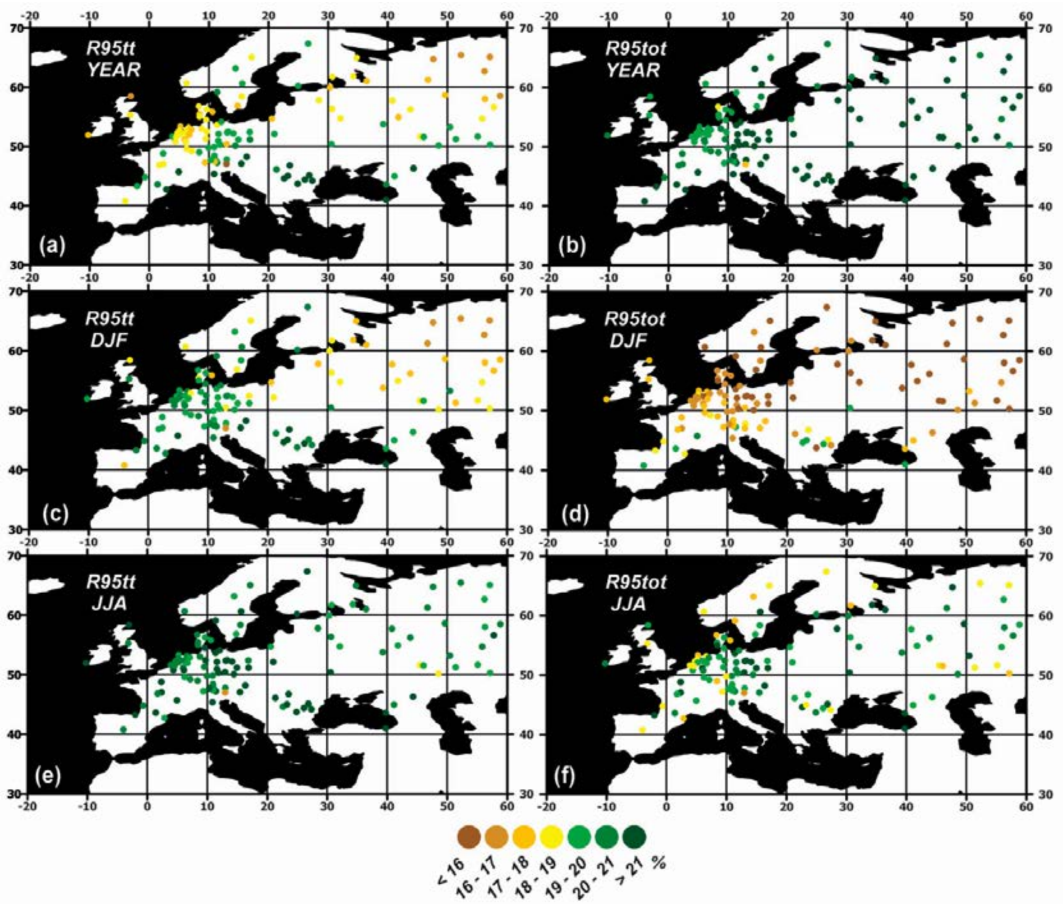

Fig. 9 *Climatological distribution (%) of (left) R95tt and (right) R95tot indices, derived from the (a, b) annual, (c, d) winter (DJF), and (e, f) summer (JJA) time series.*

The uncertainty in computing *R95tot* is not only related to the small number of wet days in a season. Even for a moderate number of wet days (e.g. 40–50 per season), there still could be a situation when none of the wet days exhibits precipitation exceeding the long-term 95th percentile. If we look at the occurrence histograms of the number of wet days for the cases when *R95tot* is nil over 51 years for all stations (roughly 25% of all cases), the distributions are peaked at 30–40 days in summer and at 40–45 days in winter (Zolina *et al.*, 2009). Thus, even when *R95tot* is above zero, its estimate may be uncertain. In other words, the uncertainty of the computation of *R95tot* results from the *finite* number of wet days rather than from the *small* number of wet days. Thus, the difference between the two indices goes far beyond just the ability to estimate relative precipitation extremeness more accurately and for shorter time intervals. The new index places this estimation in the framework of established functional distributions and links the fractional contribution from the wettest days with the initial value distribution known to be gamma distributed. Another strength of our approach is that alternatives to the gamma PDF distribution would lead to equations

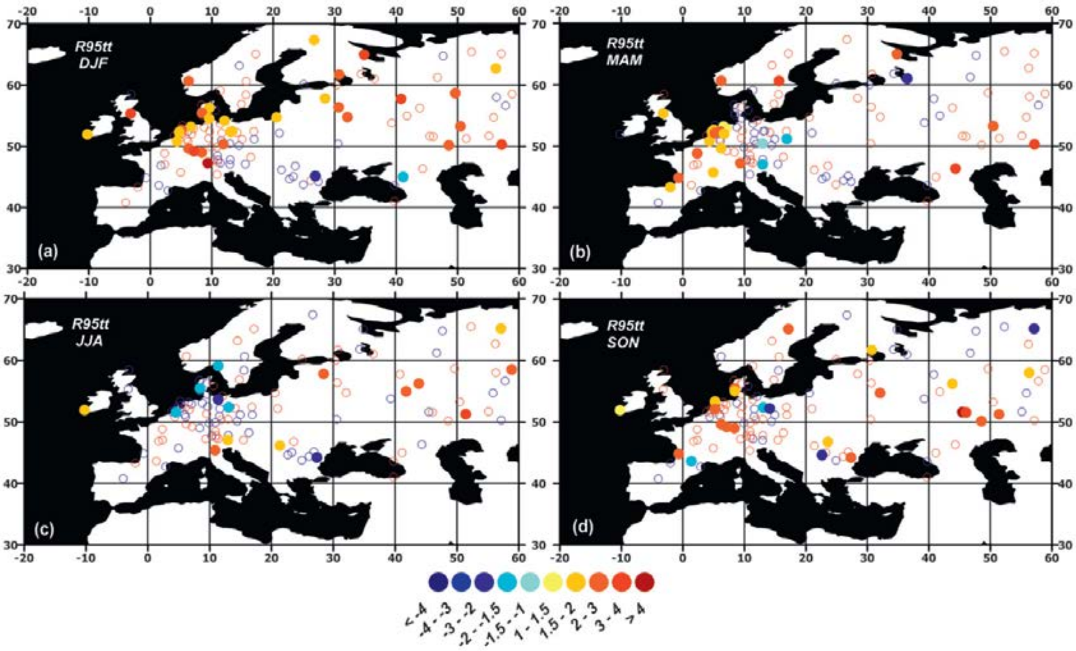

Fig. 10 Linear trends (%/decade) in R95tt for: (a) winter, (b) spring, (c) summer, and (d) autumn for the period 1950–2000. Small open circles show all trend estimates, and large circles denote the locations where the trends are significant at the 95% level. Blue symbols indicate negative trends and red symbols indicate positive trends.

similar to equations (4) and (7). In other words, the assumption of a gamma PDF is, *per se*, not critical for the derivation of DFC.

The new index *R95tt* allowed us to provide, for the first time, reliable estimates of seasonal linear trends in precipitation extremeness (Fig. 10). In winter (Fig. 10(a)), strong linear trends of up to 3–5% per decade are observed in Central Europe and in eastern European Russia. The spring pattern (Fig. 10(b)) shows a clear upward tendency (1.5 to 3% per decade) in Western Europe and in eastern European Russia, while Central Europe and northern European Russia are characterized by downward trends of from –1.5 to –2.5% per decade. The summer pattern (Fig. 10(c)) shows negative trends (as much as –3% per decade) in Western and Central Europe and positive tendencies in eastern European Russia. In autumn (Fig. 10(d)), positive trends over Western Europe are superimposed with downward trends, while in eastern European Russia the changes are positive (above 4% per decade). Overall, the trends in *R95tt* in Central Europe are upward during winter, spring and autumn and negative during summer. Over eastern and southeastern European Russia, however, trends in *R95tt* are primarily upward for all seasons.

The field significance of the positive trend pattern in *R95tt* in winter holds over the whole of Europe at the 99% level. In summer the pattern of negative trends for Western Europe is confirmed by a field significance test at the 95% level. A comparison of the seasonal trend estimates for *R95tot* (Zolina *et al.*, 2009) with those for *R95tt* reveals large discrepancies, which are most probably caused by the uncertainties in the estimation of *R95tot*. Seasonal trend patterns in *R95tot* are quite noisy. The winter-time upward tendency is confirmed by *R95tot* only in Western Europe. Noticeable discrepancies were also observed in summer for Western Europe, and during spring and autumn for Central Europe. The linear trends in *R95tt* and *R95tot* may even have different signs in some locations. *R95tot* tends to underestimate positive tendencies in winter and negative tendencies in summer, compared to *R95tt*, likely masking the seasonality signal in linear trends. Generally, it is not surprising that *R95tt* provides more robust estimates compared to *R95tot*. With *R95tt* we avoid two major uncertainties inherent in the estimation of *R95tot*: (i) constraints placed by the finite number of wet days, and (ii) uncertainties associated with the use of empirical distributions for estimation of a threshold.

R95tt and *R95tot* differ not only in their linear trends, but also in their short period interannual variability. Correlation coefficients between *R95tt* and *R95tot* derived from annual time series exceed 0.7 for 89 of 116 locations, and 0.8 for 27 locations. For the seasonal time series, correlations are considerably smaller. In winter the correlation coefficient is above 0.7 for only 63 of 116 locations.

Differences in the interannual variability of the two indices impact on the analysis of the influence of large-scale circulation patterns on heavy precipitation in Europe. Scaife *et al.* (2008) projected the winter North Atlantic Oscillation (NAO) index onto *R90tot* for the period 1900–2000. They reported significantly positive correlations with NAO in only a few locations over Scandinavia (up to 0.55) and significantly negative correlations over Southern Europe. No significant correlations were found in Eastern Europe. In Zolina *et al.* (2009) we computed correlations between the winter (DJF) NAO index (Hurrell, 1995) and heavy precipitation characteristics. For *R95tot* we obtained low positive correlations in Central Europe and Scandinavia (the highest correlation coefficient is 0.39) and negative correlations of below –0.40 over eastern European Russia and the Iberian Peninsula. With *R95tt* the level of correlation is considerably higher and the spatial pattern is more pronounced. The highest significant positive correlation with the NAO index is 0.60, with correlations in central Western Europe around 0.5. Negative correlations form pronounced patterns over Southern Europe and south European Russia with correlation coefficients varying from 0.3 to 0.5. The pattern of correlations between *R95tt* and NAO, compared to that formed by *R95tot*, is characterized by more points with significant correlations and less spatial noise. Correlation patterns of *R95tt* with NAO hold field significance at the 95% level, while a similar pattern for *R95tot* shows field significance only at the 90% level.

The new index, *R95tt*, allows us to avoid the uncertainty associated with the finite number of wet days, which is inherent in *R95tot*. This uncertainty limits the application of this measure to annual time series and to regions with sufficient numbers of wet days. Thus, *R95tt* satisfies the demand for the estimation of trends and shorter-term variability patterns on a seasonal or monthly basis. Up to now, most studies, including those which formed the basis for the IPCC AR4 (Klein Tank & Können, 2003; Groismann *et al.*, 2005; Alexander *et al.*, 2006, Trenberth *et al.*, 2007), were performed for the annual time series. The few estimates performed for seasonal time series (Moberg *et al.*, 2006; Scaife *et al.*, 2008) report patterns of trends that are highly influenced by spatial noise, likely associated with the uncertainties of estimation of heavy precipitation indices from the raw data.

6.5 CHANGES IN TEMPORAL STRUCTURE OF EUROPEAN PRECIPITATION

Besides the estimation of absolute and relative extremeness for a proper quantification of precipitation extremes, we need to account for precipitation timing and to provide the basis for the analysis of the duration of wet periods. The impact of precipitation on flooding is strongly related to the number of consecutive wet days. Catastrophic European floods (Ulbrich *et al.*, 2003) are typically associated with anomalously long precipitation periods. While such analyses are already quite developed in engineering hydrology (e.g. Madsen *et al.*, 2002), they are still in an embryonic state in climate research, even for empirical data and especially for the analysis of intensity–duration–frequency distributions. The relationship between precipitation intensity and duration allows for the estimation of PDFs for different durations and the development of intensity–duration–frequency (IDF) functions. Such IDFs, which are frequently used for engineering purposes (on minute time scales), can help to estimate return periods for precipitation events with different durations.

Figure 11 shows the monthly precipitation time series for two stations in southern Germany during July 1970, which differ considerably in their grouping despite similar total rain amounts. For Bondorf, a rather continuous precipitation period prevailed characterized by moderate precipitation, while in Dornstetten during the same period, one extremely heavy rain event was observed. Approaches to the analysis of extreme precipitation without considering timing would fail when estimating the hydrological impact of these events. Only the joint consideration of precipitation intensities and durations in the analysis of wet spells at daily time scales adequately serves this purpose.

We quantified the changes in the duration of precipitation events in the period 1950–2008 using daily data from nearly 700 European raingauges, including changes in mean and heavy intensity of precipitation in relation to the duration of events (Zolina *et*

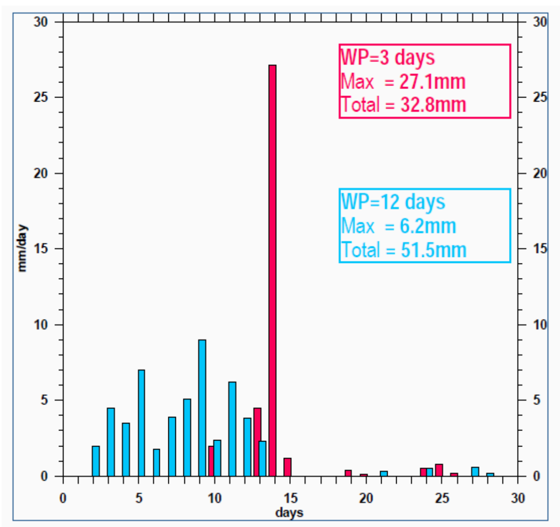

Fig. 11 *Daily precipitation time series for July 1970 for the German stations Dornstetten (red bars) and Bondorf (blue bars).*

al., 2010). For this purpose, we introduced wet spells or wet periods (WPs). For each year, WPs were quantified as consecutive days with significant precipitation (>1 mm/day). This threshold excludes very light precipitation and allows for the limited accuracy of raingauges (Klein-Tank & Können, 2003). The same threshold was used for the identification of dry day spells by Groisman & Knight (2008).

Figure 12(a) shows the overall probability of wet spell durations in Europe. Duration of wet periods decreases approximately exponentially with length. Isolated wet days (WDs) contribute about 50% to the total number of WPs. Regionally, their fraction does vary from 30 to 60%. In Zolina *et al.* (2010), we computed the normalized occurrence anomalies of WP durations during the last 60 years (Fig. 12(b)). The fractional (normalized by the total) contribution of WPs with different durations to the total number of wet days has increased. Negative anomalies of occurrence for long wet periods during the 1950s and 1960s become primarily positive in the 1990s and 2000s. The negative trend for the number of 1-day and 2-day events (Fig. 12(c)) is statistically significant, as well as the remarkably high positive trend for the occurrence of long wet periods. Importantly, this is an almost pan-European pattern.

Figure 13 shows estimates of linear trends in the duration of wet periods over Europe and clearly identifies significantly positive linear trends over central Western Europe, most parts of Scandinavia and central and northern European Russia, ranging from 2 to 4–5% per decade. Negative trends of up to 3% per decade are observed in a few locations in Southern Europe. Zolina *et al.* (2010) also proved that the lengthening of WPs is not simply caused by an increase in the total number of wet days, but by regrouping of rain events from shorter into longer WPs. We evaluated the net effect of the observed increase of wet days over Europe (regionally up to 3.6% per decade) on the trends in the duration of WPs using a Monte Carlo approach. Three groups of experiments were performed. Firstly, the observed change in the number of wet days was simulated by random generation of single wet days. This may also result in a decreasing duration of WPs due to the increasing frequency of single-day events. Secondly, the event generation was performed according to the observed frequency distribution of WPs (Fig. 12(a)). Finally, in an "extreme" simulation the observed

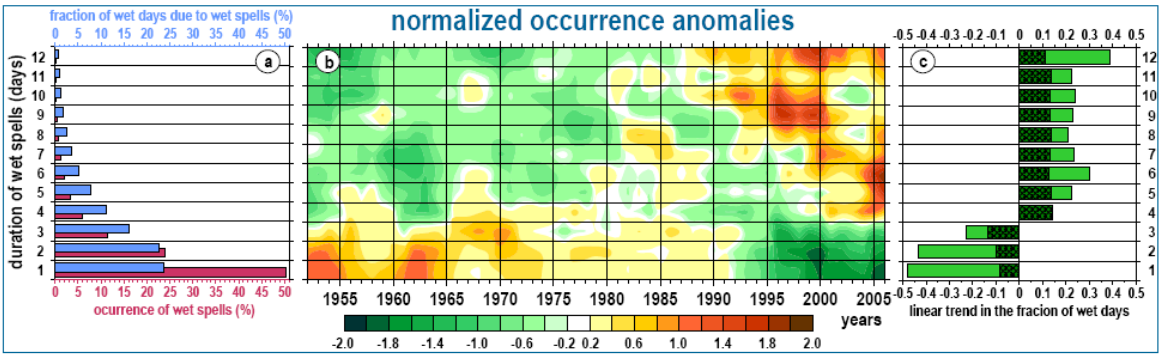

Fig. 12 *(a) Climatological distributions of the occurrence of WP durations (red) and of the fractional contribution of different WPs to the total number of WDs (blue); (b) temporal evolution of the normalized occurrence anomalies of the contribution of different WPs to the total number of WDs for all European stations smoothed with a 5-year RM; and (c) linear trends (% per decade) in the occurrence of WPs of different duration (green bars) along with their 95% significance (dark shaded bars).*

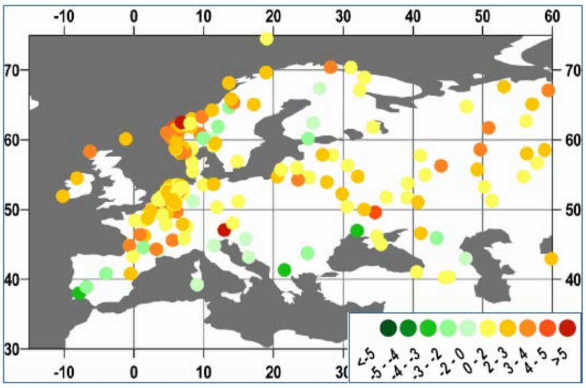

Fig. 13 Linear trends (% per decade) in the mean duration of wet periods over Europe (1950–2008).

change in the number of wet days was generated by one continuous WP. The results show that the maximum expected net effect from the change of the number of wet days, even for the "extreme" unrealistic case, may impose trends in the duration of WPs of only 0.02–0.16 days over a 60-year period. This is four to nine times smaller than the observed changes in wet spell durations.

Analysis of wet spells was further extended to consideration of dry spells by application of theoretically derived statistical distributions of the duration of wet and dry periods. Theoretically, the durations of wet/dry spells are distributed according to the geometric distribution. Although it is formally close to the exponential, the geometric distribution is a discrete distribution whose variables can take on only discrete values (Spiegel, 1992). Since the duration of wet and dry spells from daily data can only be estimated in discrete days, an approach employing discrete distributions is much better justified compared to that based on life-time distributions. The choice of geometric distribution within the family of discrete distributions is not critical and some other distributions (e.g. log-Series, LS; Mixed Geometric, MG) can be considered. Practical analysis (e.g. Deni *et al.*, 2010) shows that the results from the most discrete distributions are close to each other and start to be different for extremely long wet and dry spells (e.g. in monsoon regions). More important, in building the geometric distribution, we need to account for the finite number of wet days, because the largest wet/dry period can be only equal to the total number of wet/dry days. In Zolina *et al.* (2011), we used the truncated geometric PDF given by:

$$P(x_i = k) = \frac{1}{1-(1-p)^N}\, p(1-p)^{k-1} \tag{6}$$

where p is the distribution parameter, k the duration in days in the wet/dry period, k = 1, 2, … N, and N is the number of wet/dry days, implying the length of the *theoretically* longest wet/dry period. Using this distribution we obtained, in Zolina *et al.* (2011), robust estimates of the tendencies in the duration of wet and dry periods over Europe for different seasons. In particular, estimation of the tendencies in the duration of extremely long wet and dry periods for the cold and warm seasons (Fig. 14) clearly demonstrated that, at least in Central and Eastern Europe during the last several decades, we clearly observe the regrouping of the wet days into prolonged wet and dry episodes.

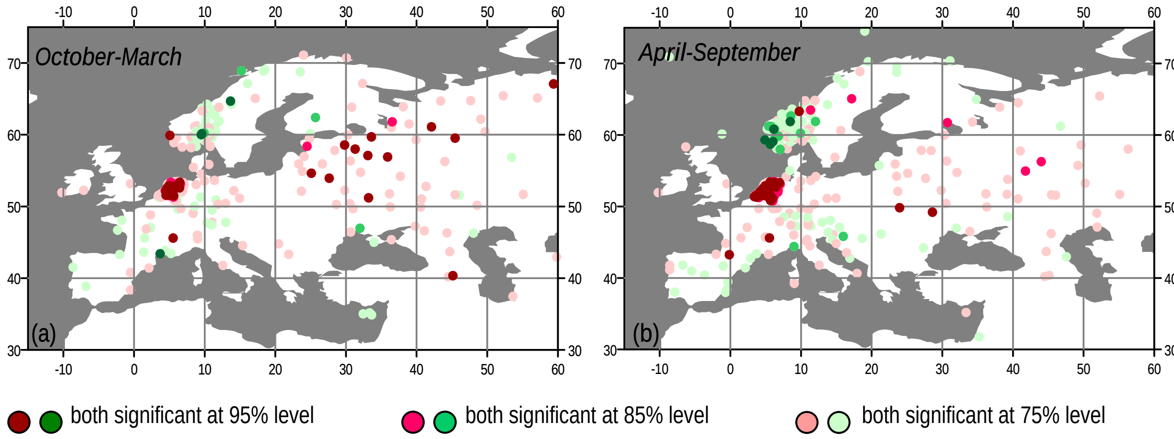

***Fig. 14** Locations for which linear trends (over 1950–2010) in the duration of extremely long (90th percentile) wet and dry periods show the same sign and are statistically significant at different levels for: (a) cold and (b) warm seasons. Red circles correspond to the positive trends and green ones indicate negative trends.*

This indirectly confirms also the conclusions of Zolina *et al*. (2010) about a minor role of the change in the number of wet days in the lengthening of wet spells. Alternatively, the durations of dry spells had to be shortened to compensate for the effect of the growing number of wet days on the length of wet episodes.

It is also important to note that in becoming longer, wet periods over Europe are also characterized by more abundant precipitation. Heavy and extreme precipitation events during the last two decades have become much more frequently associated with longer wet spells and also become more intense in comparison with 1950s and 1960s. During 1950–2008 the occurrence of the association of extreme rainfall with longer WPs has increased by 3–4% per decade, while the occurrence of extreme events during shorter WPs (1–2 days) has decreased by 3% per decade. The percentage of extremes associated with WPs shorter than 3 days decreased from 60% in the 1950s–1960s to 45% in the 1990s–2000s, while the occurrence of extremes associated with longer WPs increased from 40% to 55%. Furthermore, the extremes associated with longer WPs intensified over Europe with upward trends in intensity of 2–3% per decade in Western Europe and >5% per decade in European Russia, implying actual changes of 4–9 mm/day over the 60-year period. At the same time, extremes associated with shorter WPs either do not show any trend, or locally decrease, e.g. in central Western Europe, by 2–4% per decade.

These findings allowed us to extend the joint analysis of durations and intensities of precipitation events to climate time scales. Figure 15(a) and (b) shows trends in the probabilities of occurrence on the intensity–duration diagram for the mean intensities (a) and for extreme precipitation quantified by the 95th percentile (b). The occurrence of wet periods longer than 2–3 days with high mean daily precipitation intensities has clearly increased, while short and moderately long wet periods with light precipitation (less than 4 mm/day) have decreased in frequency. Thus, wet spells providing relatively large precipitation totals (per period) have increased, while short wet periods providing small totals have decreased. The intensity of precipitation extremes

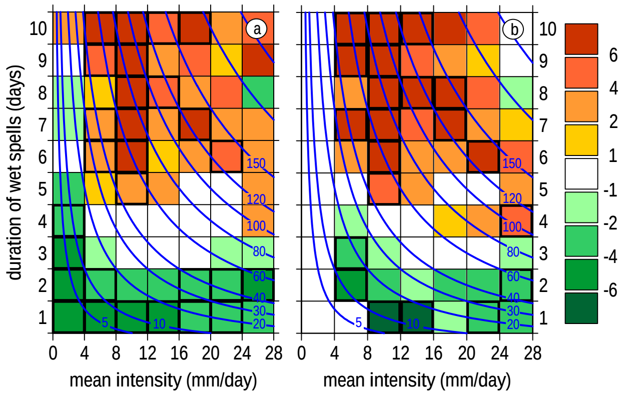

Fig. 15 *Trends (% per decade) (a) in the frequency of WPs of different duration and mean intensities and (b) in the frequency of WPs associated with extreme precipitation. Bold-sided squares mark trends significant at the 95% level. Blue contours show precipitation totals for WPs of different durations and mean intensities.*

(exceeding the 95th percentile) occurring during long wet spells increased by 6–8% per decade, while extremes occurring during short wet periods weakened by 4–6% per decade. Thus, indeed, longer European wet spells result in more abundant precipitation.

The lengthening of the European wet spells combined with an increased occurrence of extreme events hints at an increasing role of moisture advection by mid-latitude cyclones in forming precipitation extremes with a pace exceeding that implied by local temperature changes (Trenberth *et al.*, 2003; Allan & Soden, 2008). The direct association of cyclone activity with the duration of WPs requires the analysis for individual seasons, which is difficult due to the limited number of wet spells per season or month. In this respect, an accurate derivation of the theoretical PDFs for the WP duration and associated precipitation intensity should provide good prospects for further analysis.

6.6 SUMMARY AND CONCLUSIONS

This overview of climate variability of European precipitation extremes shows that heavier and extreme precipitation in Europe exhibited complex variability over the last five decades. Absolute precipitation extremes increased in winter with trend magnitudes of 4 to 10% per decade (about 1–3 mm/day per decade). The strongest change is observed in Eastern Europe. Although in summer an upward tendency is also prevalent, it is weaker than for the winter season in Eastern Europe and can change to the opposite tendency in Western Europe. Regionally, changes in heavy and extreme precipitation in Central Europe exhibit clear seasonality. The 95th and 99th percentiles of daily precipitation increase from 5 to 13% per decade in winter, spring and autumn and decrease by 3 to 9% per decade during summer. Estimates of field significance show that the seasonality found is insensitive to including or excluding the first and the last several years of the records.

Relative extremeness of precipitation on average follows long-term tendencies in the absolute extremes. However, regionally, patterns of linear trends in relative extremeness are quite noisy. This can be partly explained by the uncertainties of estimation of extreme precipitation indices based on the raw data. To overcome this problem, we suggested a new index for extremeness based on the distribution of the fractional contribution of very wet days to total precipitation. The new index *R95tt* is more stable than *R95tot*, especially when precipitation extremes are estimated from the limited number of wet days of seasonal and monthly time series. The new index allows for an accurate estimation of trends in extremeness for individual seasons, and quantification of seasonality in relative extremeness. When annual daily time series are analysed, linear trends in *R95tt* and *R95tot* are qualitatively consistent; both hint at a growing occurrence of extreme precipitation of up to 3% per decade in central Western Europe and in south European Russia, with a somewhat more evident trend pattern for the *R95tt* index. Linear trends estimated for individual seasons, however, exhibit pronounced differences when derived from both indices. In particular, in winter, *R95tt* clearly reveals an increasing occurrence of extreme precipitation in western European Russia (up to 4% per decade), while during summer, a downward tendency in the fractional contribution of very wet days is found in central Western Europe. The new index also allows for a better association of European extreme precipitation with the North Atlantic Oscillation (NAO) index by showing a more consistent spatial correlation pattern and higher correlation levels compared to *R95tot*.

European precipitation has not only increased and become more extreme during the last 60 years, but also its structure has changed. Short and isolated rain events have been regrouped into prolonged wet spells. Extreme precipitation events associated with longer WPs have intensified by about 12–18% during the last 60 years, while extremes associated with short WPs became less intense. Simultaneously, the duration of dry spells has also increased over Europe, showing that this effect is not modulated by the changing number of wet days but rather associated with the re-grouping of rainy events. From the hydrological perspective, it is important to account for the potential impacts of the changed precipitation regime on water resources and flooding. While there is no significant trend in flood frequencies for the past 80–150 years in Europe (Mudelsee *et al.*, 2003), a clustering of floods has occurred in the last decades (Bunde *et al.*, 2005). The changing character of WPs may significantly influence change in the frequency and strength of floods. Lengthening of WPs will also enhance groundwater recharge and soil moisture storage (Maxwell & Kollet, 2008). However, it is not easy to evaluate the transient hydrodynamic response at the catchment scale without integrated hydrological simulations, accurately reproducing the variability across all spatio-temporal scales from days to decades and from metres to kilometres.

Besides the impacts of extreme precipitation on European flooding, an important question is: what are the reasons for the observed changes in the intensity of precipitation extremes and precipitation temporal structure? Trenberth *et al.* (2003) argued for the importance of the advective mechanisms as the main factor leading to changes in the intensity of heavy precipitation. In winter, excessive lateral advection of moisture can be likely associated with mid-latitudinal cyclones. Given the increasing intensity of precipitation extremes and a tendency of wet periods towards lengthening, we can hypothesize that the growing frequency of very deep cyclones (Loeptien *et al.*, 2008) over Europe, along with the increasing occurrence of cyclone series (Mailer *et al.*, 2006) may play a leading role, at least in winter. These two phenomena are likely to be intensified in the future climate (Loeptien *et al.*, 2008, Ulbrich *et al.*, 2009). At the same time, in summer there are other potential mechanisms that may result in intensification of precipitation extremes, such as local moisture balance that is largely constrained by the regional orography. To investigate these mechanisms in more detail, extensive numerical experimentation, preferably with high-resolution non-hydrostatic modelling, is required. In this respect, development of new generation regional re-analyses based on advanced model formulations may shed more light on our understanding of the mechanisms of changing precipitation extremes in Europe.

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