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Large scale analysis of changing frequencies of rain-on-snow events and their impact on floods

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Abstract

In January 2011 a rain-on-snow (RoS) event caused floods in the major river basins in Central Europe, i.e. the Rhine, Danube, Weser, Elbe, Oder, Ems basins. This event prompted the question how to define a RoS event and whether those events have become more frequent. Based on the flood of January 2011 and on other known events of the past, threshold values for RoS events were determined and consequently events which combine average rainfall of at least 3 mm on a snowpack of at least 10 mm snow water equivalent (SWE) as well as 20 % melt content by summing rainfall and snowmelt were analysed. RoS events were estimated for the last 61 yr and for the entire study area based on a temperature-index snow model driven with the European-scale E-OBS data. Frequencies and magnitude differ depending on the elevation range. When distinguishing alpine, upland, and lowland basins, we found that upland basins are most influenced by RoS events. Over the last decades their occurrences shifted from late to early winter. Overall, the frequency of rainfall increased in the winter, while the frequency of snowfall decreased in the spring. In nearly all lowland and upland basins an increasing trend in the frequency of RoS events since 1980 was observed. These results suggest an increasing flood hazard from RoS events in January and February in the medium mountain ranges of Central Europe, especially in the Rhine, Weser, and Elbe river basins.

1 Introduction

Rain-on-snow (RoS) events are relevant for water resources management, and especially for flood forecast and flood risk management (McCabe et al., 2007), since they can cause flood events in the winter season. These events are of major interest because they depend not only on the rain intensity and amount, but also on the prevailing freezing level, the snow water equivalent (SWE), the snow energy content, the timing of release, and the areal extent of the snowpack (Kattelmann, 1997; McCabe et al., 2007). Snowpacks are water reservoirs of large regional extent and storage capacity, which can produce rapid melt in combination with warm air temperatures and high humidity (e.g. Singh et al., 1997; Marks et al., 1998). Consequently, cumulating rainfall and snowmelt can increase the magnitude of runoff and can have much greater potential for flooding than a usual melt event (Kattelmann, 1985; Marks et al., 1998). The scientific interest for this kind of events was raised in the last decades by global warming and several methods of analysis and measurement were developed to understand and quantify the physical processes for basins in different climatic locations and elevations (e.g. Blöschl et al., 1990; Singh et al., 1997; Floyd and Weiler, 2008) on individual events. Many studies observed an increase in the occurrence of liquid precipitation in the winter time and an earlier snowmelt due to increase of air temperature in Europe and the USA (e.g. Birsan et al., 2005; Hamlet et al., 2005; Knowles et al., 2006; Renard et al., 2008). Furthermore, Köplin et al. (2013) predicted a shift from snowmelt dominated runoff catchments to a more variable snow and rain-fed regime in the future in Switzerland. Their simulations also suggest a diversification of flood types in the winter time and an increase of rain-on-snow flood events in the future (Köplin et al., 2013).

Only few studies specifically analysed the changes of the frequency of RoS events over time and especially over large areas. Ye (2008) observed that the increase in air temperature resulted in an increase of rainfall days in winter in Northern Eurasia. In a further study, Ye et al. (2008) were able to correlate the increase in RoS days with the increase in air temperature for the same study area. Sui and Koehler (2001) studied the RoS induced flood events in Southern Germany using long-term (1961–1995) observed data at 10 meteorologic stations and 17 hydrologic gauging stations in a forested region of the Northern Danube tributaries at elevation ranges from 320 to 1456 m a.s.l. They found decreasing trends in SWE for nearly all stations, which could be explained by temperature trends, and increasing trends in maximum daily winter precipitation. Even though the precipitation in winter is less than in summer in this area, Sui and Koehler (2001) observed larger extreme values of peak discharge in winter than in summer, which led them to the conclusion that the water volume through snowmelt plays a key role in RoS events. They also observed a positive trend in the winter flood quantiles at all gauging stations. Furthermore McCabe et al. (2007) analysed monthly RoS events for the years 1949 through 2003 in the western USA. Their results showed generally positive temporal trends of RoS events frequencies for high elevation sites and negative trends for low elevation sites. In these areas, the increase of temperature seems to affect the occurrence of snow, contributing therefore to the lower frequency of RoS events (McCabe et al., 2007). Finally, Surfleet and Tullos (2013) analysed the frequency of RoS events in the Santiam River basin in western United States for three elevation bands using Global Circulation Models. Their results showed that an increase in air temperature due to climate change would lead to a decrease of high peak flow due to RoS events for low and middle elevation zones, while at high elevation bands, this kind of events would increase.

Most studies differ on the characterisation of RoS driven events. McCabe et al. (2007) and Surfleet and Tullos (2013) defined an event as RoS driven, if simultaneously liquid precipitation occurs, maximum daily temperature is greater than 0 °C, and a decrease in snowpack can be observed, while for Ye et al. (2008), a RoS event takes place only when at least one of the four daily precipitation measurements is liquid and the ground is covered by 1 cm snow or more. Sui and Koehler (2001) found that most RoS events in Southern Germany occurred when snowmelt was larger than the rainfall depth. These definitions allow identifying all possible RoS events but are insufficient if one focusses on the events that can effectively cause flood events.

A good example of such events is the RoS event that occurred in January 2011 in Central Europe. Due to a strong negative phase of the North Atlantic Oscillations, the temperature anomalies in December 2010 reached −4 °C in Middle and North Europe (Lefebvre and Becker, 2011), and led to especially large amounts of snow all over Germany, with snowpacks reaching the maximum observed snow depth since beginning of the measurements at some locations (e.g. Böhm et al., 2011; LHW, 2011; Besler, 2011). January 2011 brought thawing temperatures in combination with rainfall events and from 6 to 16 January very high water levels were observed at nearly all German gauging stations (e.g. Böhm et al., 2011; Bastian et al., 2011; Karuse, 2011; LHW, 2011; Fell, 2011; Besler, 2011). Kohn et al. (2013) clearly identified the simultaneous occurrence of liquid precipitation and snowmelt as the driving factor for those flood events, that led, beside other impacts, to a restriction of navigation on the Rhine and large inundations in the lower Elbe basin.

Due to the great hydrologic impact that RoS events can have, there is a real need for assessing the changes in frequencies of RoS events and for identifying, which events may cause large flooding. The aims of this study are therefore (i) to characterise RoS driven events leading to potential extreme flooding in Germany, using the case study of January 2011, and (ii) to analyse the changes in frequencies and magnitudes of this type of events for six major Central European basins, i.e. Rhine, Danube, Elbe, Weser, Oder and Ems.

2 Materials and methods

2.1 Study area

The study area embodies the six major river basins of the German fluvial network. Since only German streamflow records were used, the basins of the rivers Rhine, Danube, Elbe, Weser, Oder and Ems are considered only upstream of the most downstream station on German territory (Fig. 1). According to the Hydrological Atlas of Germany (HAD, Bundesanstalt für Gewässerkunde), the basins were divided into alpine, upland, or lowland sub-basins, depending on their main elevation range.

Only a small part of South Germany is located in the alpine mountains, this area spreads from Allgeau, through part of Bavaria, towards Salzburg and reaches its highest point at the Zugspitze (2963 m a.s.l.). The Rhine and Inn (Danube) origin in the Swiss an Austrian Alps and Alpine regime can therefore play an important role in those basins.

From the southern border of Germany to the southern boundary of the North German lowlands, as well as in the northern part of the Czech Republic (Weser basin), the landscape can be described as upland with elevation ranges from 200–300 m a.s.l. to up to Feldberg, 1493 m a.s.l., the top of the Black Forest in Baden-Württemberg.

North Germany and West Poland (Oder basin) are mainly constituted of lowland areas with altitudes ranging from 0 to 200 m a.s.l.

2.2 Meteorological and hydrometric data

Daily mean temperature and precipitation sums were obtained from the European Climate Assessment and Dataset from the ENSEMBLES project (Haylock et al., 2008). The so-called E-OBS dataset (version 6.0) was interpolated from climate stations all across Europe into a $0.25^{\circ} \times 0.25^{\circ}$ regular latitude longitude grid (Fig. 1). The time series are available from 1 January 1950 to 31 December 2011 and cover the study area: 46.00–55.25° N and 5.25–19.75° E.

Daily mean discharge data from more than 300 gauging stations in Germany were provided by German public authorities. Details on the assembled dataset can be found in Kohn et al. (2013). The time series are of different lengths but most of them cover the period 1950–2011. All data included in this analysis passed a visual quality control for suitability in this study, for which some minor corrections to the peaks that may be done a couple of years later as it is routine correction practice after floods by the authorities will be irrelevant. The location of the gauging stations used for the analysis can be found in Fig. 2.

2.3 Estimation of the snowpack and the snowmelt

Snow accumulation and melt were estimated based on daily E-OBS mean temperature and precipitation sum data for the entire study area and are given in mm SWE. Precipitation is assumed to be solid when air temperature $T_a < 1\,^{\circ}\mathrm{C}$ and liquid when $T_a \geq 1\,^{\circ}\mathrm{C}$. Snowmelt M (mm) is estimated using a temperature-index-model, which assumes a relationship between ablation and air temperature (Eq. 1; e.g. Finsterwalder and Schunk, 1887; Collins, 1934; Corps of Engineers, 1956).

$$M = M_\mathrm{f} \cdot (T_\mathrm{a} - T_\mathrm{b}) \tag{1}$$

Despite its apparent simplicity compared to energy balance methods, Ohmura (2001) showed that the temperature-based melt-index model was sufficiently accurate for most practical purposes, and was justified on physical grounds since the air temperature is the main heat source for the atmospheric longwave radiation. The advantage is that it needs only daily precipitation and temperature data and is therefore optimal for large scale analysis. Martinec and Rango (1986) calculated degree-day factors M_f for open areas depending on the snow density of the snowpack. They suggested M_f values from 3.5 to 6 mm °C^{-1} day^{-1} and even smaller for fresh snow. They also observed that M_f increases over the melt period. Hock (2003) listed M_f values for snow in high elevation areas between 2.5 and 6 mm °C^{-1} day^{-1}. For sake of simplicity of the large scale analysis, a constant conservative value of $M_\mathrm{f} = 3$ mm °C^{-1} day^{-1} was chosen for the entire study area and melt period. This value was found to represent the area well, since snow melts very fast in upland and lowland regions in Germany and the snowpack consists therefore mainly of fresh snow. The base temperature T_b represents the threshold temperature for melting snow. Most studies set T_b to 1 °C, since energy is needed to bring the snow to 0 °C to start melting (e.g. Hock, 2003). T_b was therefore set to 1 °C. Before, however, the sensitivity of the subsequent trend calculation to the choice of T_b was tested ranging from 0 to 2 °C and found to be rather insensitive.

For every cell, daily snowpack SP is calculated for day i as the sum of the SP of the day before and the snowmelt (mm) or solid precipitation P_S (mm SWE) of the actual day (Eq. 2) and is given in mm SWE.

$$\mathrm{SP}_i = \mathrm{SP}_{i-1} \begin{array}{ll} +P_{S,i}, & \text{if } T_\mathrm{a} < T_\mathrm{b} \\ -M_i, & \text{if } T_\mathrm{a} \geq T_\mathrm{b} \end{array} \tag{2}$$

Snowpack was estimated for a period from 2 to 1 August of the following year. At the beginning of each hydrologic year SP was set to zero. Therefore it only accounts for annual snow and no multi-year storage is taken into account in the alpine basins. The results show the winter period from 1 November to 31 May taking into account potential snow season extents in the entire study area.

2.4 Characterisation of RoS events

The flooding potential of a RoS event can be determined by comparing the equivalent precipitation depth, defined as the daily sum of snowmelt and liquid precipitation, to the corresponding discharge measurement. The return periods of discharges were calculated with the generalised extreme value distribution and the parameters were estimated with the maximum likelihood method (Venables and Ripley, 2002).

To compare the equivalent precipitation depth with discharge, 12 gauging stations were selected (Fig. 1) at the outlet of each sub-basin. The discharge of a sub-basin corresponds to the difference between the discharges measured at the outlet stations of the sub-basin and of the upper sub-basin. No discharge data were available for the Oder basin outside the German borders, therefore only the station at the outlet of the lowland sub-basin was considered. Since the upland part is very small compared to lowland, the influence of the upland regime is assumed to be minor for the type of events considered in the Oder basin.

Based on the case study of the large scale RoS driven flooding event in January 2011 in Germany and Central Europe (see Sect. 3.1 and Kohn et al., 2013), we identified the driving parameters for an event to be RoS driven and defined threshold values, that were validated on historic events (Table 1). Those parameters are:

1. Snowpack SP: Since the interest is on events potentially leading to great floods and driven by rain and snowmelt, the SWE should be large enough to significantly contribute to runoff.

2. Snowmelt *M*: Generally, the amount of *M* in the equivalent precipitation depth must be substantial to define an event as RoS driven.

3. Rainfall P_L: In a RoS event, snowmelt is enhanced by liquid precipitation. The amount of rainfall must also be substantial. Otherwise the event may be only snowmelt driven.

The flooding time and magnitude depend on the response time of a basin. The discharge corresponding to a given RoS event was defined as the peak discharge occurring between the day *i* when all threshold values (see above) are reached and the following days until the response time is reached. The equivalent precipitation depth of a RoS event corresponds then to the sum of the daily precipitation and snowmelt from the first day of melting before the day *i* until the day of occurrence of the maximum discharge. If snow continues to melt and if there is still liquid precipitation between two RoS events and the two corresponding discharges, the event is considered as one event with two or more flood waves. This avoids the overlapping of two events in a time series. The term RoS event will further be used only for events as defined above, otherwise, the term RoS day will be used, corresponding to a day when all the thresholds for rain-on-snow (see above) were reached.

2.5 Trend analysis

Equivalent precipitation depth, as well as the other parameters influencing the occurrence of RoS events, namely snowpack, rainfall and snowfall, were analysed for trends for all alpine, upland, and lowland sub-basins.

Trends were calculated using a linear regression and are given in % of the expected value. They represent the positive or negative slope of the linear function, for increasing or decreasing trends, relative to the mean value of the time series. This allows comparing the importance of the changes in time of a parameter in one basin with the others. They were tested at a 5 % significance level using the non-parametric Mann–Kendall statistic test (Mann, 1945). Trends were calculated and compared for the time periods 1950–2011 and 1990–2011, corresponding in this paper to the long-term and short-term trends.

3 Results

3.1 Rain-on-snow event in January 2011

Figure 2a–f shows the mean basin wide daily snowpack (mm SWE), snowmelt (mm), and rainfall (mm), as well as the percentage of snow covered cells in the Rhine, Danube, Elbe, Weser, Ems, and Oder river basins in January 2011. On 5 January, the snowpack was spread over 100 % of the cells in the study area and the mean SWE varied from 25 mm (Ems) to 70 mm (Danube). The overall flooding of January 2011 can be described as a two waves flood event corresponding very well to two phases of rainfall combined to snowmelt. The first flood wave (W1), from 6 to 10 January, was caused by rainfall and snowmelt spreading from West to East. It generated a total snowmelt between 34 and 47 mm for a total rainfall between 17 and 24 mm in the western basins Rhine, Weser, and Ems. In the eastern basins Danube, Elbe, and Oder a total snowmelt from 19 to 29 mm and a total rainfall from 3 to 9 mm were observed. During this event, daily liquid precipitation of 2 to 10 mm day^{-1} fell on the western basins (Fig. 2a, c, d), while in the eastern basins it never exceeded 2 mm day^{-1} (Fig. 2b, d, f). The daily equivalent precipitation depth reached 10 to 26 mm day^{-1} in the West and 2 to 14 mm day^{-1} in the East. In both regions the percentage of snowmelt in runoff was higher than 25 %. In the western basins both rain and snowmelt played therefore an important role, while in the eastern basins the first flood wave was mostly snowmelt-driven. During the second wave (W2), from 11 to 16 January, a total snowmelt of between 2 and 23 mm and an average total rainfall of 25 mm could be observed in the western basins. In the eastern basins, the event generated an average total snowmelt of 35 mm for a total rainfall from 14 to 26 mm. The rainfall reached 0 to 10 mm day^{-1} in the western part and 2 to 17 mm day^{-1} in the eastern part. Equivalent precipitation depths between 0 and 18 mm day^{-1} were observed in the West and between 5 and 28 mm day^{-1} in the East. On 10 January, most of the snow had already melted in the northwestern part of Germany and the event was therefore rather rain-driven in this area, especially in the Ems river basin. However, the snowpack was still substantial at the time in Central Germany, and the RoS event caused annual maximum daily runoff with up to 100 yr return periods in the upper parts of the Elbe, Weser, and Rhine basins (Fig. 2, map). After W2, nearly all snow was melted in the Ems, Weser and Oder basins. In the Rhine river basin, most of the remaining snow was located in the alpine region. In the Danube basin, the snowpack was still substantial, but also mainly in the alpine region.

Comparing the cumulated equivalent precipitation depth at different elevation zones Fig. 2 shows that the alpine zone was only little impacted with an average potential runoff of 4 mm during W1 and W2. The alpine sub-basin of the Danube was more affected by the RoS event of W2 than by the melt-driven event of W1. Upland areas, especially in the West, reacted very strongly to the RoS events leading to equivalent precipitation depth in all basins of almost 25 mm day^{-1} during W1 and in the Weser again during W2. After the events, nearly all the cells were free of snow. In the East, W1 equivalent precipitation depths were mostly due to snowmelt for all elevation ranges and showed very similar reaction for all lowland and upland sub-basins, with average equivalent precipitation depths of 10 mm day^{-1} for Elbe, 3 mm day^{-1} for Oder and 8 mm day^{-1} for Danube. W2, induced in the East by the cumulation of rain and snowmelt, led to a very fast increase of the equivalent precipitation depths within few days especially for the upland sub-basins of Elbe and Danube. The Weser and Ems lowland areas, in the western half of Germany, were also strongly impacted during W1, since the amount of snow was substantial and unusual, and it nearly completely melted by the end of W1 and W2 therefore had little impact. The Rhine basin usually has a lot of snow during the winter time, due to its larger upland elevation ranges and the influence of its alpine part. For this reason, there were not many differences in the snowmelt processes of the Rhine lowland and upland sub-basins in January 2011, which were strongly impacted by both RoS events.

3.2 Characterisation of a rain-on-snow event

As observed during the RoS driven flooding event in January 2011 in Central Europe (Sect. 3.1), average basin wide SWE reached at least 10 mm before W1 and W2. The rainfall depth was at least 2 mm and the snowmelt was at least 25 % of the equivalent precipitation depth. Therefore threshold values of 10 mm SWE for snowpack, 20 % for snowmelt amount in the equivalent precipitation depth, and 3 mm for rainfall were chosen to define RoS days. The maximum duration of the flood wave in the January 2011 event was 6 days and was used as the basin response time for the selection of the RoS events. The duration of an event can therefore vary from one to dozens of days. Figure 3 compares the equivalent precipitation depth and the corresponding discharge of the RoS events selected using the above threshold values to all possible RoS events (SP > 0 mm SWE, $M > 0$ %, $P_L \geq 0$ mm) for all river basins and elevation ranges. Historic events, for which RoS processes were identified in the literature as the main cause for flooding, are listed in Table 1. The past RoS events were well caught by all defined thresholds and the amount of selected events is reasonable. In the Elbe lowland, no event was identified for January 2011. This is due to the fact that the RoS event mainly impacted the Elbe upland, which can be seen by comparing the snowmelt in upland and in lowland areas during the first flood wave from 6 to 10 January (Fig. 2). The second snowmelt events did not impact the lowland since snow was already melted.

For every basin, RoS events are of different importance. The alpine and upland basins of Rhine, Weser and Danube showed the highest positive correlation of the discharge to the sum of equivalent precipitation depth during the RoS events with correlation coefficients between 0.68 and 0.75 (Fig. 3). This means that discharge peaks in those basins can be strongly influenced by RoS events. In the other upland basins Elbe and Ems, correlation coefficients of 0.48 and 0.51 were found and in lowland none of the correlation coefficients was higher than 0.48, which means that in these areas, discharge is little influenced by the equivalent precipitation depth and rain plays the more important role.

The percentage of RoS events leading to a discharge of a given quantile in a month is represented in Fig. 4. In the alpine Rhine and Danube, all RoS events led to quantiles of 0.7 to 1 for every month except May. Most of the events corresponded to discharge peaks with quantile 0.9–1 for all winter months, with progressively increasing amount from March to May. In upland basins, events occurred from December to April, with the most events leading to discharge of 0.9–1 quantiles from January to March. Only few events corresponded to discharges of quantiles < 0.7 confirming the strong flood-causing potential. Lowland areas were more heterogeneous in their reaction to RoS events. They occurred only from December to March and had corresponding discharges of quantiles from 0 to 1, which amplifies the observation in Fig. 3, that RoS events do not necessarily cause the highest floods. Very heterogeneous are the quantiles of RoS events for the months January–March in the lowland basins of Rhine, Oder, and Elbe, and most of the events occurred in March. In the Weser and Ems lowland basins, the RoS events are very infrequent but the few events that occurred between December and March led mostly to relatively high discharge peaks (quantiles 0.8–1).

3.3 Trends

Figure 5 shows the yearly sum of equivalent precipitation depth of all RoS events according to the thresholds described in Sect. 3.2 in the entire winter, in the early winter (November–February), and in the late winter (March–May) from winter 1950/51 to 2010/11. The trends were calculated only for years with RoS events and represent therefore the change in the magnitude of the events. In the alpine basins Rhine and Danube, RoS events generated a potential runoff for the entire winter between 100 and 600 mm year^{-1}, and corresponded on average to 45, 22, and 72 % of the total winter, early winter and late winter precipitation (liquid and solid). In both basins the equivalent precipitation depth was greater in late winter than in early winter. No clear trends were identified in the magnitude of the equivalent precipitation depth in the early winter, but late winter showed for both basins decreasing magnitudes, leading to an overall decreasing trend for the entire winter.

Upland basins generated maximum equivalent precipitation depths from 90 mm year^{-1} in the Ems basin to up to 400 mm year^{-1} in the Danube basin (Fig. 5c–g). The equivalent precipitation depth corresponded to an average of 21 %, 28 %, and 35 % of the total winter, early winter and late winter precipitation. In those regions, RoS events occurred more frequently in the early winter than in the late winter. In the Rhine basin (Fig. 5c), the positive trend in the late winter was emphasized by the large RoS event that occurred in the 1980s and the events are becoming less frequent during the second half of the time period. In contrary, even if the late winter did not show any trend in the magnitude of the equivalent precipitation depth, the events in this period have become more frequent. The same changes in frequencies of RoS events were observed for the Elbe and Weser river basins (Fig. 5e and f), leading to the conclusion that in these upland basins, the occurrence of RoS events shifted from late winter to early winter from the first to the second half of the time series. RoS events occurred more often in the early winter than in the late winter in the Danube upland basins and both seasons showed decreasing trends (Fig. 5d). In the Ems upland basin, RoS events occurred very seldom and mostly only in the early winter. A decreasing trend of those events was observed (Fig. 5g).

In the lowland basins, RoS events were rather rare and generated maximum equivalent precipitation depths from 70 mm year^{-1} in the Oder basin to up to 250 mm year^{-1} in the Rhine basin (Fig. 5h–l), corresponding to an average of 13, 18, and 29 % of the total winter, early winter and late winter precipitation. Since the occurrence of RoS events is infrequent, they depend on very specific climatic conditions and can occur either in the early winter or in the late winter. The Rhine lowland showed the greatest amount of RoS events in the winter time, which is due to the fact that a small part of the basin contributing to the runoff in this area consists of medium elevation mountain ranges. As for the upland areas, the frequency of the

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events shifted from late to early winter for the second half of the time series in the Rhine, Elbe, and Weser lowland basins (Fig. 5h, j, and k). For all lowland basins except the Oder, the magnitude of the events decreased in late winter. In early winter, the magnitude increased in Rhine, Weser, and Ems basins and decreased in Oder and Elbe basins.

In Fig. 6 long-term trends of the total equivalent precipitation depths of all possible RoS days ($\mathrm{SP} > 0\,\mathrm{mm\,SWE}$, $M > 0\,\%$, $P_\mathrm{L} \geq 0\,\mathrm{mm}$) and of the selected ones ($\mathrm{SP} > 10\,\mathrm{mm\,SWE}$, $M > 20\,\%$, $P_\mathrm{L} \geq 3\,\mathrm{mm}$), the average snow water equivalent, as well as the solid and liquid precipitation are shown as % yearly change relative to the mean of the period for the monthly sum from November to May and for the seasonal sum over the winter for all basins. In contrary to Fig. 5, the trends were calculated including years without RoS events and thus also account for changes in the frequency of occurrence. Significant trends at $p < 0.05$ are shown with a red star. Around 10 % of the trends in the equivalent precipitation depth of all events and 2 % of the trends in the selected events were significant. For mean SWE 42 % of the trends were significant, for snowfall around 30 %, and for rainfall 20 %. Overall, the detected long-term trends were rather small, ranging between −4 and +4 % yearly change relative to the mean of the period. In the upland basins the trends were positive in January and February and negative from March to May. In the lowland basins the trends were negative for all winter months. The trends of equivalent precipitation depth for the selected RoS days were very similar to the ones for all RoS days in alpine and upland basins. Lowland trends, in contrary, differed from all RoS days with positive trends from November to January. SWE decreased for all elevation ranges and all winter months, with especially large negative trends in April in some upland and all lowland basins. As well as SWE, the number of days with $\mathrm{SP} > 10\,\mathrm{mm\,SWE}$ also showed decreasing trends between January and April in the upland and lowland basins, especially in April in the Ems and Weser basins, but no trends were identified in the alpine regions. As snow water equivalent, snowfall also showed overall decreasing trends. In contrary, rainfall increased in the winter time from

January to March in upland and lowland and in November, December, and April in the alpine basins.

The short-term trends (Fig. 7) showed, compared to the long-term ones, stronger negative and positive trends ranging from less than −10 % to more than +10 %. For all parameters, a maximum of 20 % of the trends were significant. Trends in the equivalent precipitation depth of all RoS days were positive in the alpine basins for all winter months. The trends were negative in November, December, and April and positive from January to March in upland and lowland basins. A very strong increase in the equivalent precipitation depth of the selected RoS days could be observed from January to March in upland and lowland basins. SWE showed high positive short-term trends from December to March and negative ones in May in the upland and lowland basins, opposite to the long-term trends, which had negative trends for all winter months. The snowfall trends were also increasing from November to March and decreasing in April. Rainfall magnitude as well as the occurrence of days with a liquid precipitation sum of at least 3 mm increased in February and May and decreased in the rest of the winter for all basins.

The sensitivity analysis of the base temperature showed that increasing or decreasing T_b by ±1 °C had only a small impact on the direction of the trends. On average for long-term and short-term trends 95 % of trend values showed the same direction. The equivalent precipitation depth of the selected RoS days was the most sensitive with around 80 % of the trends having the same direction. 100 % of the rainfall trends showed the same direction.

4 Discussion

The RoS driven flood event that spread over Germany and Central Europe in January 2011 is a good example of the very large scale impact that RoS events can have on the hydrologic cycle. From 6 to 16 January nearly all gauging stations reached the maximum water level observed in the year. The large scale impact of this

RoS event was on the one hand due to the extremely broad snowpack, that covered mostly the entire study area at the beginning of January, as well as its remarkable depth, that reached extreme values at several locations in Germany (e.g. Böhm et al., 2011; Bastian et al., 2011; Karuse, 2011; LHW, 2011; Fell, 2011; Besler, 2011). The snowpack represented therefore a very large water reservoir available for runoff. On the other hand, the climatic conditions, leading to thawing temperatures and rainfall in January, provided the required energy for melting the snowpack. The runoff was therefore generated in a very short time, approximately 6 days for each flood wave, and reached simultaneously the upper and lower parts of the Rhine, Danube, Weser, Elbe, Oder and Ems streams leading to an overall increase of the water levels and to the maximum annual discharge in most areas. Even if most of these discharges, corresponding to return periods of less than 10 yr, were not statistically extreme (Kohn et al., 2013), this RoS event emphasizes the large scale impact of such events and their potential of shifting the annual peak flow from spring to early winter. It represents therefore a good reference for the characterisation of RoS events with large risk of flooding.

Using this RoS driven flood events as a reference, it was possible to identify the snow depth, the percentage of snowmelt in the equivalent precipitation depth, and the magnitude of rainfall as the main characteristics of a RoS event and as the major parameters for the runoff generation. The resulting threshold values of 10 mm SWE for snow depth, 3 mm for rainfall, and 20 % for snowmelt allowed us to keep most historic events in the time series and to eliminate the insignificant ones from the time series. They are therefore good indicators for RoS driven flood events. The snowpack threshold (10 mm SWE) is not only representative for the event of January 2011, but also corresponds to the definition of the beginning of the winter given by Beniston (2012) and Bavay et al. (2013). The advantage of identifying RoS events with threshold values is that it can easily be applied to other basins, where discharge data are available, and it represents a useful tool for analysing the changes in frequencies and magnitudes of those events.

The results showed that RoS events have generally a large impact on discharge peaks in alpine and upland basins, most likely occurring from January to March and most likely leading to discharges of monthly quantiles 0.7–1 in the Alps and even 0.8–1 in upland. This is in agreement with the observation of Sui and Koehler (2001) that RoS events play a more important role in runoff generation than pure rainfall events for topographical elevations larger than about 400 m a.s.l. In all lowland basins and Elbe upland, the quantile ranges are in contrary strongly influenced by the pre-conditioning of the stream in the winter time, which is due to the fact that discharge in those areas is rather rain-fed in the winter, since snowfall occurs only infrequent. In those areas, a RoS event is likely to have a small impact on discharge if the discharge is low before the event or in contrary a large impact if the discharge is already high. Therefore RoS events in those areas do not necessarily cause floods but they can exacerbate a flood. For example, the Elbe river basin generated discharges corresponding to return periods of up to 100 yr during the January 2011 event not only because of the RoS event, but also because of the very wet autumn 2010, which led to already very high water levels and discharges at the beginning of January (e.g. Kohn et al., 2013).

The trend analysis of the estimated magnitude of the RoS events showed a shift in the occurrence from late winter to early winter in the upland and lowland regions. This can be explained by the decreasing trends in snow depth observed in spring in these areas (Figs. 6 and 7) and therefore the decreasing probability of rain falling on a snow covered soil in spring. In contrary, the trends in rainfall are positive in the early winter, increasing the probability for RoS events, especially in January and February. RoS events also became more frequent since the 1980s, especially in upland and lowland regions. These results correlate well with the observations of many studies worldwide (e.g. Birsan et al., 2005; Knowles et al., 2006; Ye, 2008; Ye et al., 2008).

Trends have to be interpreted carefully since they depend on the choice of the time period, which can be influenced by many climatic factors and also extreme values. The analysis of long-term (1950–2011) vs. short-term (1990–2011) trends showed different, even opposite, results for all parameters. Overall, long-term trends are smaller than short-term trends. The greatest difference by comparing both periods is in the mean snow water equivalent. In the long-term analysis, snowpack is receding for all basins, while it increases in the short-term trends for all basins but the Alps. Fricke (2006) observed that the average snowpack in Germany for the period 1961–1990 was substantially higher than for the period 1991–2000. The 21st century was characterised by extreme events. The winter 2005/06 and December 2010, for example, were identified as extremely snow rich in Germany (Fricke, 2006; Pinto et al., 2007; Böhm et al., 2011). This explains the differences in the trends, since the extreme events of the 21st century will take more importance in a shorter time series, especially if at the beginning of the period rather small snow depth were observed. Alpine basins show different trends than upland and lowland basins. While upland and lowland have negative trends in the long-term trend analysis and positive in the short-term, snow water equivalent decreased in the Alps for both, long- and short-term trend analysis. This can be explained by the climatic conditions specific to the alpine regions, which are very different than in the upland and lowland basins. For example, while exceptionally great SWE were measured all over Germany in December 2010, the Swiss Alps experienced snowfalls below average (e.g. Trachte et al., 2012; Techel and Pielmeier, 2013). In another study in the Swiss Alps, Beniston (2012) found out that the winter time precipitation declined between 15 and 25 % over the 1931–2010 period for almost all stations. He found out that the number of snow-sparse winters has increased in the last 40 yr, while the number of snow-abundant winters has declined. But in the meantime, some winters since the 1990s have encountered record breaking in snow amount and duration (Beniston, 2012). Rainfall also shows opposite trends, increasing for the long time series and rather decreasing for the short ones in the winter time. Snowpack and rainfall both influence the occurrence of RoS events and trends in RoS days are therefore difficult to identify for the long-term analysis, since rainfall is increasing but snow depth is decreasing. The results showed rather positive trends for upland and negative ones for lowland. In the short-term analysis in contrary, clear positive trends are detected for all RoS days in upland and lowland regions. The trends for the selected

RoS days are even more positive, leading to the conclusion that they have become an important factor for the winter discharge and that the occurrence of maximum peak flow between January and March is getting more frequent since the 1990s.

One challenge in the trend analysis of extreme RoS events is the censored data, i.e. events do not occur every year. Most methods for trend analysis try to statistically disclose outliers, as for example the Sen-slope method. In our case, the outliers are often exactly the values that need to be analysed. Therefore, the linear regression was the most reliable method for the trend analysis. The zeros also explain why many trends were not significant. This is a well-known problem in hydrology. Kundzewicz et al. (2012) observed for example that the strong natural variability of hydrologic events can alter trend detection and IPCC (2012) pointed out that, due to the fact that extreme events are per definition rare, most of the record length are not long enough to allow for detection of trends in a time series. Therefore, trends cannot be considered as absolute values. However, the sensitivity analysis showed that there were no significant differences in the direction of the trends for all parameters, which is a good indication that the trends are reliable.

The runoff generation from RoS events is influenced by many physical processes as the thermal, mass, and wetness conditions of the snowpack, or the snow metamorphism, the water movement through the wet snow, the interaction of melt water with underlying soil, or the overland flow at the snow base (Singh et al., 1997; Marks et al., 1998). For the calculation of the actual runoff, evaporation, infiltration, or sublimation processes should be as well taken into account and a physically based model for the estimation of snowpack and snowmelt would be needed. In the case of a large scale analysis, it is difficult to assess and predict all parameters and all processes, the challenge is therefore to accurately estimate the runoff with the only available data, namely temperature and precipitation. The conceptual temperature-index model employed in this study allowed to estimate the potential snow depth and snowmelt. Even if the degree-day factor was generalised for the entire study area, the method led to a good estimation of the potential runoff and was accurate enough to recognise known historic events. This validation of RoS events and the selection and analysis of potential flood generating RoS events add to previous studies, which have mostly looked only at RoS days.

5 Conclusions

In a context of climate change, changes in extreme events are very likely (IPCC, 2012). Snowpack and precipitation forms in the winter time are very likely to change and therefore may influence the occurrence and magnitude of rain-on-snow events. Defining threshold values to characterise RoS events allowed to identify the events with potentially high impact on discharge at the large scale and to analyse them for trends in frequencies and magnitude. The results showed an elevation dependence of RoS events, which have the strongest impacts in upland regions. An increasing frequency of these events could be observed since 1980 in the upland basins. In lowland and upland basins, RoS events occurred in the first half of the period 1950–2011 mostly in the late winter and in the second half rather in the early winter. This shift correlates well with the decreasing trends in snowpack in spring and the increasing trends in rainfall in early winter.

The results showed the importance of the choice of the analysed period for the trend analysis, since opposite trends were found for snow water equivalent in the long-term and short-term analysis. The 21st century seems to be characterised by several extreme events which makes the analysis even more difficult. As the example of January 2011 in Germany and Central Europe showed, rain can release a large amount of water stored in the snowpack and RoS events can cause very wide-spread flood events, bringing a large amount of water into the streams within a very short time. Since these events are very likely to become more frequent in the future in certain basins and elevation ranges and therefore the flood hazard increases in the winter, there is a real need for an improved understanding of the relation between the RoS events and flooding and more analysis is needed on their occurrence at different scales.

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Table 1. Historic RoS events (sources: Wetterchronik, 2001; Kohn et al., 2013).

Date	Province	Basin	Event description
18 Mar 1970	Lower Saxony	Elbe	Snowmelt and rainfall led to flooding all over the river, especially in Uelzen and Lüneburg.
27–28 Feb 1987	Bavaria	Danube	Large floodings due to snowmelt and incessant rainfalls in Schambachtal.
20 Jan 1997	Rhineland-Palatinate	Rhine	Small floodings after a snow rich January and a very wet February.
6–10 Jan 2011	Central Europe	Western basins: Rhine, Weser, Ems	First wave of large scale floodings due to snow rich December 2010 followed by thawing temperatures in January 2011.
11–16 Jan 2011	Central Europe	Generalised over all basins, especially in the East	Second wave of large scale floodings due to snow rich December 2010 followed by thawing temperatures in January 2011.

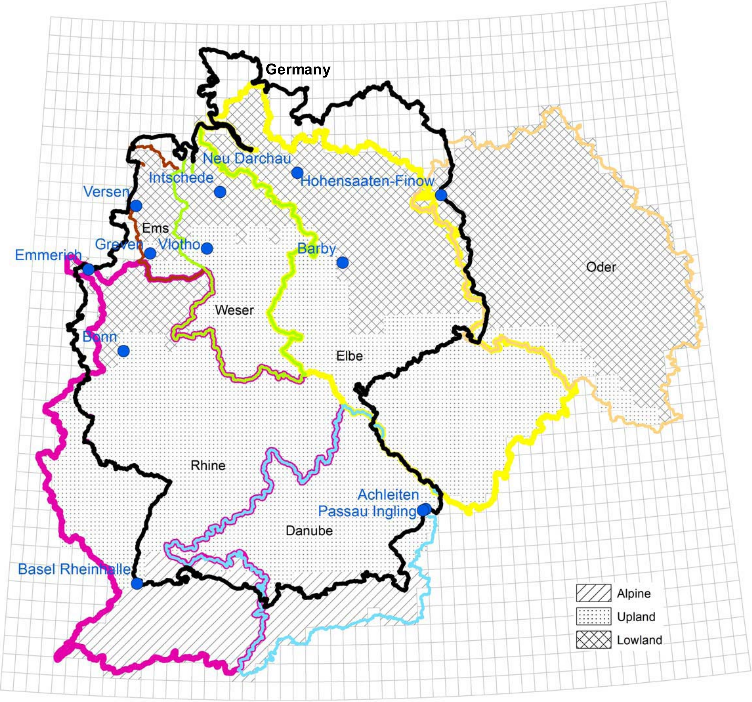

Fig. 1. Study area: delimitation of the basin boundaries of the Rhine, Danube, Elbe, Weser, Oder, and Ems basins and subdivision in alpine, upland, and lowland. The background raster corresponds to the E-OBS dataset. Gauging stations at the outlet of each sub-basin are represented with a blue dots.

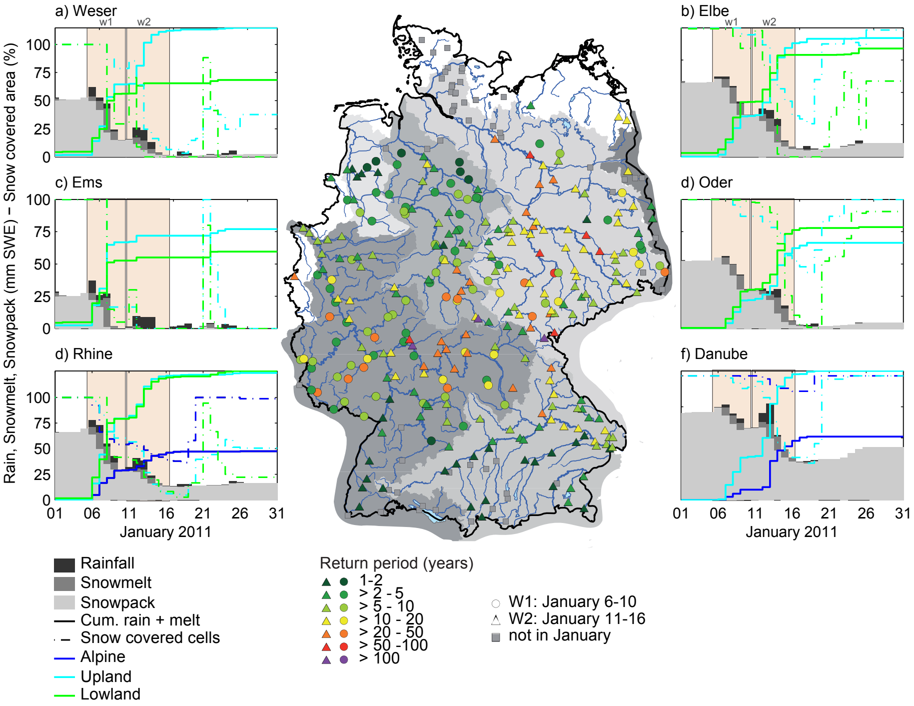

Fig. 2. Large scale analysis of the RoS event of January 2011. **(a–f)** Mean basin wide daily snowpack, liquid precipitation, and snowmelt (mm SWE) from 1 to 31 January, as well as the percentage of snow covered cells. Map: return period and occurrence period of the maximum daily discharge in the calender year 2011 at all gauging stations (modified from Kohn et al., 2013).

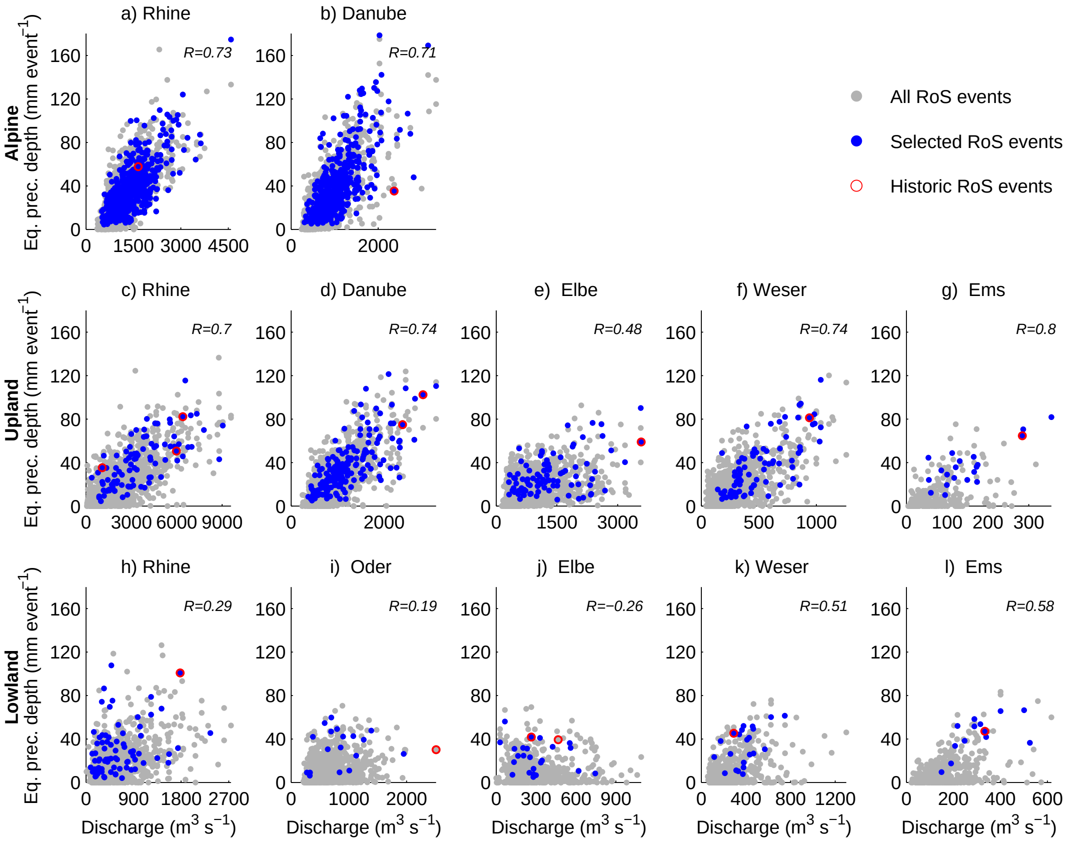

Fig. 3. Total equivalent precipitation depth and corresponding discharge for all possible RoS events (SP > 0 mm SWE, $M > 0\,\%$, $P_L \geq 0$ mm), for all selected events (SP > 10 mm SWE, $M > 20\,\%$, $P_L \geq 3$ mm), and for documented historic events. The correlation coefficient R is given for the selected RoS events.

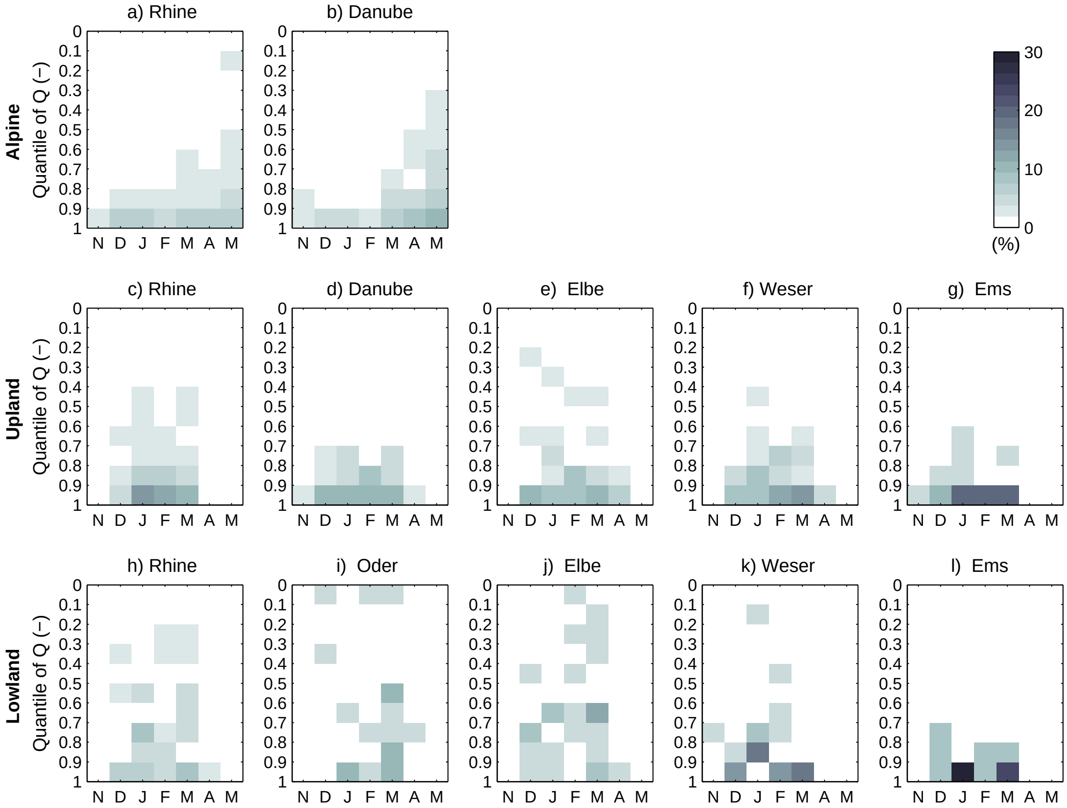

Fig. 4. Percentage of all selected RoS events from 1950 to 2011 by month of occurrence (November–May) and discharge quantile.

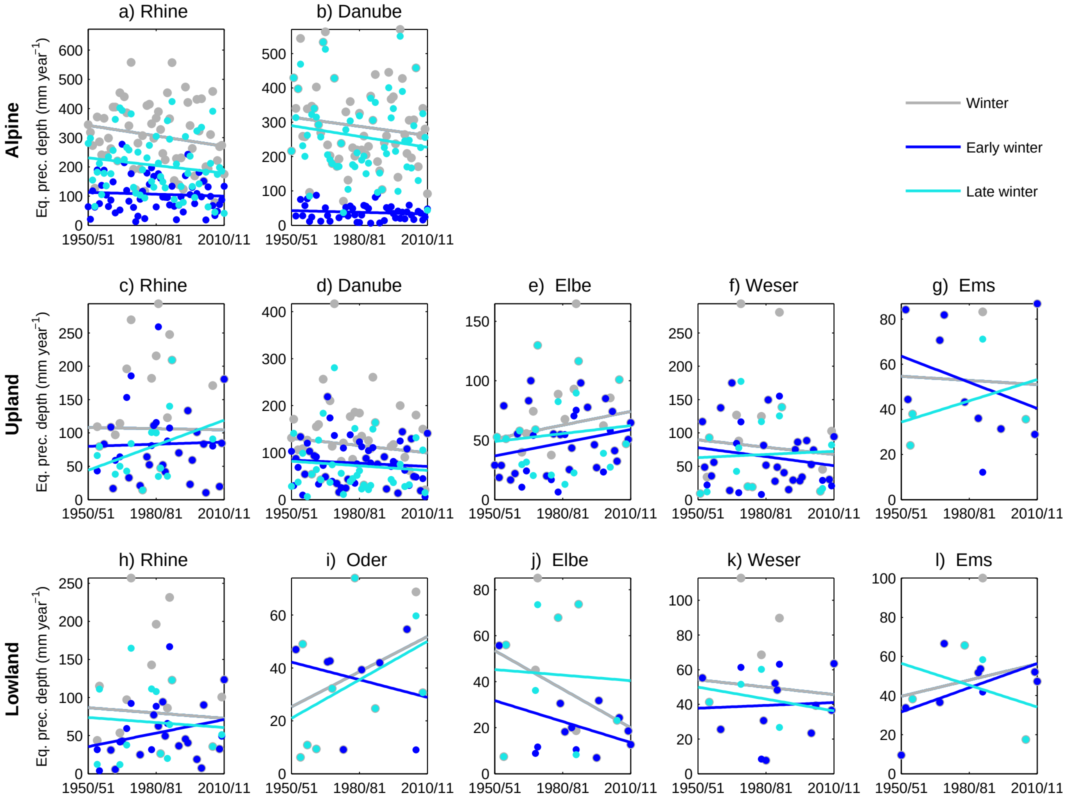

Fig. 5. Total equivalent precipitation depth for all selected RoS events in the winter (November–May), in the early winter (November–February), and in the late winter (March–May) for the period 1950–2011. Only the years with RoS events are represented.

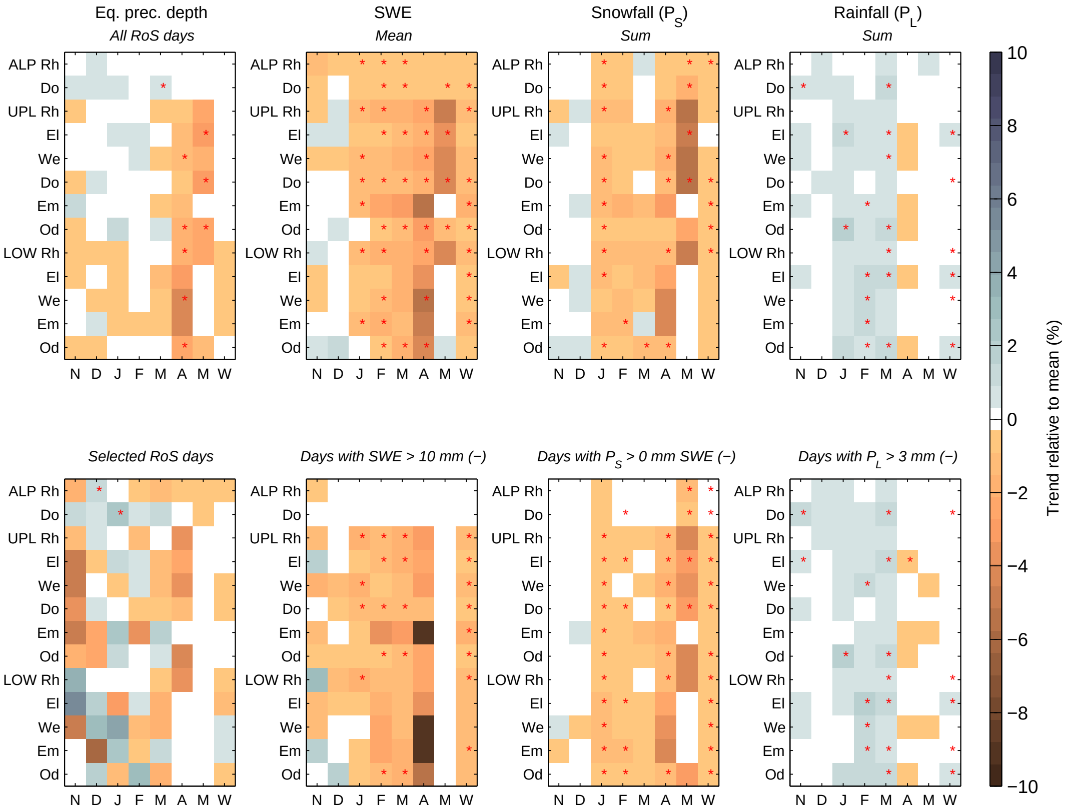

Fig. 6. Comparison of the longterm trends (1950–2011) of the total equivalent precipitation depth, the snow water equivalent, the snowfall and the rainfall for the months November–May (N–M) and the entire winter (W) for the alpine (ALP), upland (UPL) and Lowland (LOW) subbasins of Rhine (Rh), Danube (Do), Elbe (El), Weser (We), Oder (Od), and Ems (Em). Trends are given as the yearly change relative to the mean of the period 1950–2011. The significant trends at $p < 0.5$ are shown with a red star.

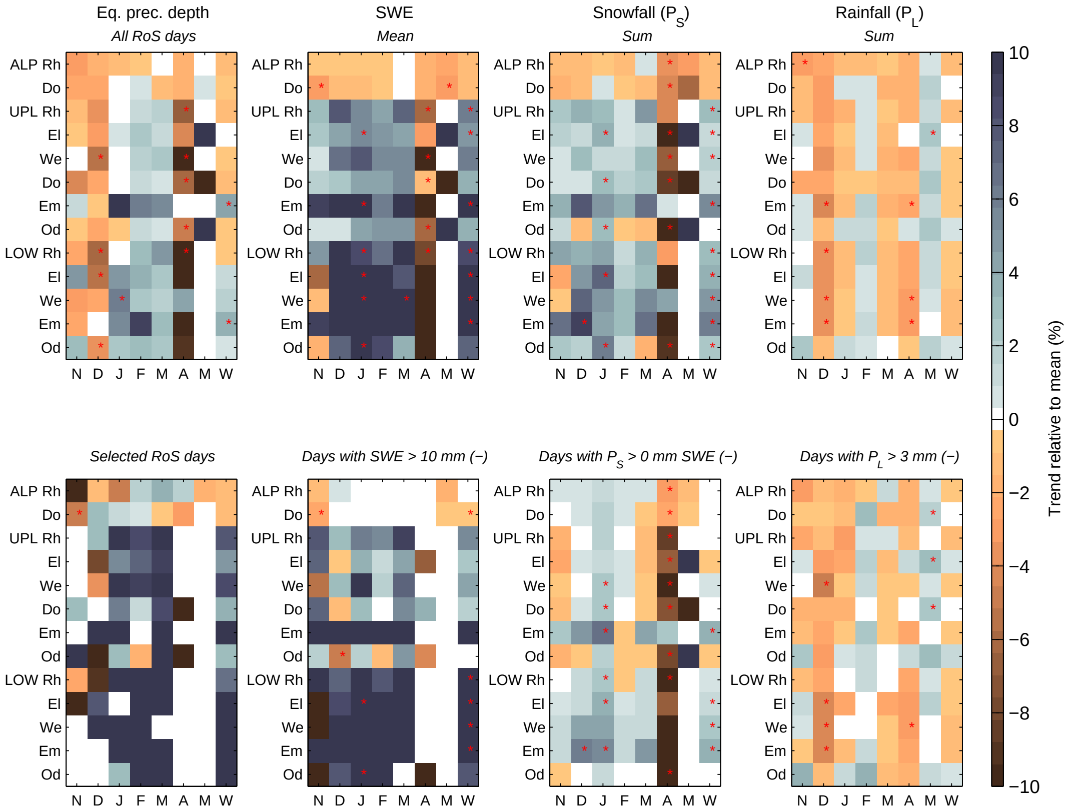

Fig. 7. Comparison of the shortterm trends (1990–2011) of the total equivalent precipitation depth, the snow water equivalent, the snowfall and the rainfall for the months November–May (N–M) and the entire winter (W) for the alpine (ALP), upland (UPL) and Lowland (LOW) subbasins of Rhine (Rh), Danube (Do), Elbe (El), Weser (We), Oder (Od), and Ems (Em). Trends are given as the yearly change relative to the mean of the period 1990–2011. The significant trends at $p < 0.5$ are shown with a red star.

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