Theor. Appl. Climatol. (2009) 95: 9–26
DOI 10.1007/s00704-007-0361-x
Printed in The Netherlands

Theoretical and Applied Climatology

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Winter precipitation trends for two selected European regions over the last 500 years and their possible dynamical background

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With 8 Figures

Received 28 April 2007; Accepted 12 October 2007; Published online 11 April 2008

Summary

We analyse winter (DJF) precipitation over the last 500 years on trends using a spatially and temporally highly resolved gridded multi-proxy reconstruction over European land areas. The trends are detected applying trend matrices, and the significance is assessed with the Mann–Kendall-trend test. Results are presented for southwestern Norway and southern Spain/northern Morocco, two regions that show high reconstruction skill over the entire period. The absolute trend values found in the second part of the 20th century are unprecedented over the last 500 years in both regions. During the period 1715–1765, the precipitation trends were most pronounced in southwestern Norway as well as southern Spain/northern Morocco, with first a distinct negative trend followed by a positive countertrend of similar strength. Relating the precipitation time series to variations of the North Atlantic Oscillation Index (NAOI) and the solar irradiance using running correlations revealed a couple of instationarities. Nevertheless, it appears that the NAO is responsible in both regions for most of the significant winter precipitation trends during the earlier centuries as well as during recent decades. Some of the significant winter precipitation trends over southwestern Norway and southern Spain/northern Morocco might be related to changes in the solar irradiance.

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1. Introduction

Precipitation is one of the most relevant climate elements affecting human economies and terrestrial ecosystems (e.g. Groisman and Legates 1995; Schmidli et al. 2002; Knippertz et al. 2003; Xoplaki et al. 2004; Zveryaev 2004; Santos et al. 2006; Gimmi et al. 2007; Pauling and Paeth 2007). Knowledge of the amplitude of precipitation variability, linked with events such as floods, droughts, or even glacier fluctuations (see Nesje et al. 2001, 2007; Nesje and Dahl 2003), at regional and local scale over the past centuries is important from a societal as well as scientific point of view (Le Quesne et al. 2006; Pauling et al. 2006; Esper et al. 2007; Solomon et al. 2007). To assess future climate change, enhanced understanding of the mechanisms of natural climate variability on different spatio-temporal scales for the last millennium is necessary (Pfister 1999; New et al. 2001; Nesje and Dahl 2003; Luterbacher et al. 2004, 2006; Xoplaki et al. 2004; Casty et al. 2005; Touchan et al. 2005; Esper et al. 2007).

Trends are an inherent feature in the climate system (e.g. Houghton et al. 2001; Hannachi 2007). In climatology, a trend is usually defined as a long-term change of climate elements, i.e. a shift in the average climatic conditions (Rapp 2000). The quantification of the magnitude of climatic trends and their attribution to naturally and anthropogenically enhanced climate variability have recently become a growing research field (e.g. Rapp 2000; Giorgi 2002; Hegerl et al. 2003; Stott 2003; Dose and Menzel 2004; Hunt and Elliott 2006; Zhang et al. 2007). As mentioned by Rapp (2000) and Hunt and Elliott (2006), a strong trend is indicative of a short duration, thus highlighting the danger of extrapolating such trends into the future. González-Rouco et al. (2000) and Rapp (2000) describe the importance of trend analyses as a basis to verify simulated/reconstructed historical climate variations.

In this paper, we analyse the recent high-resolution reconstruction of European precipitation over the past 500 years by Pauling et al. (2006) on winter (DJF) trends. A trend detection method hitherto not applied on data obtained from climate reconstructions is used here. This paper is structured as follows: The datasets used and the methods are described in Sect. 2. Results from southwestern Norway and southern Spain/northern Morocco are presented in Sect. 3, along with a link to the dynamical interpretation of the significant trends found. Concluding remarks are drawn in Sect. 4.

1.1 Precipitation reconstructions and trend analyses

Since precipitation is spatially and temporally highly variable, a great density of stations is needed to properly capture precipitation patterns (New et al. 2000; Schmidli et al. 2002; Knippertz et al. 2003; Trigo et al. 2004; Mitchell and Jones 2005; Efthymiadis et al. 2006; Pauling et al. 2006; Solomon et al. 2007). Long instrumental precipitation measurements along with a high network density are, however, only available over the last few decades to centuries (Groisman and Legates 1995; Hurrell et al. 2004; Moberg et al. 2006; Gimmi et al. 2007 and references therein). To assess precipitation variability further back in time, information from indirect sources (so-called "proxies") is needed. These include precipitation indices based on documentary evidence (see Brázdil et al. 2005 for a review) and various precipitation sensitive proxies from natural sources (see Sect. 2).

Continuous precipitation reconstructions in Europe covering the last few centuries are only available for a limited number of areas (Luterbacher et al. 2006; Pauling et al. 2006; Gimmi et al. 2007). The seasonally resolved reconstruction of European precipitation over the last 500 years by Pauling et al. (2006) used in this study is unique in its spatial extent and temporal resolution.

For single precipitation time series, many trend analyses exist: e.g. Osborn et al. (2000) calculated trends of daily precipitation records during the 20th century for the UK. Moberg et al. (2006) assessed trends for selected daily station records in Europe over the period 1901–2000. However, only a few studies investigate trends based on gridded precipitation data: Schmidli et al. (2002) performed linear trend analyses on monthly resolved precipitation during the 20th century for the European Alps. Efthymiadis et al. (2006) calculated trends of monthly precipitation totals covering the 1800–2003 period for the Greater Alpine Region. Giorgi (2002) presented an analysis of summer and winter precipitation trends throughout the 20th century over 22 land regions of sub-continental scale. Hunt and Elliott (2006) used a model simulation running over the last 10,000 years to detect trends of different climatic variables during selected periods. Therefore, our study using the European precipitation reconstruction by Pauling et al. (2006) is a first attempt to detect long-term trends, i.e. $\geq$30 years, over the past 500 years.

1.2 Dynamic background

Beside the detection of winter precipitation trends in the Pauling et al. (2006) reconstruction, we also investigate possible causes of the significant trends. These may include internal oscillations (primarily the North Atlantic Oscillation [NAO] and the El Niño/Southern Oscillation [ENSO]) or external forcings (changes in solar irradiance and volcanic eruptions: see Robock 2000; Hegerl et al. 2003; Yoshimori et al. 2005; Bengtsson et al. 2006; Stendel et al. 2006; Fischer et al. 2007). We focus on the NAO and solar irradiance, although the ENSO as the globally dominating mode of interannual climate variability (e.g. New et al. 2001; Rodríguez-Puebla et al. 2001;

Brönnimann et al. 2007), or strong tropical volcanic eruptions by perturbing the radiative balance of the Earth through the emission of ash particles into the stratosphere (e.g. Robock 2000; Ammann and Naveau 2003; Fischer et al. 2007) may also cause precipitation trends. Compared to the NAO and solar forcing, the ENSO is less stationary (e.g. Raible et al. 2005; Brönnimann et al. 2007), and the volcanic forcing only persists for 1–3 years (e.g. Robock 2000; Hegerl et al. 2003; Yoshimori et al. 2005; Fischer et al. 2007), which is too short to leave an impact on the long-term precipitation trends investigated in this study. Even a cumulation of strong eruptions was not found to have an important effect on the trends (not shown).

1.2.1 North Atlantic Oscillation

The NAO is the dominant pattern of atmospheric circulation variability in the North Atlantic region during winter (e.g. Hurrell 1995; Hurrell and van Loon 1997; Wanner et al. 2001; Luterbacher et al. 2002). During NAO positive (negative) conditions, the pressure systems over Iceland and the Azores are well (weakly) developed and, thus, the westerlies are strengthened (weakened), i.e. the NAO Index (NAOI), defined as the standardized pressure difference between the Azores and Iceland, is positive (negative). While southwestern Europe, parts of central Europe, and parts of the western Mediterranean experience dry (wet) conditions during the positive (negative) phase of the NAO (e.g. Hurrell and van Loon 1997; Xoplaki et al. 2004; Lohmann et al. 2005), the opposite is found for northern Europe. According to Dai et al. (1997), the NAO accounts for $\sim$10% of the 1900–1988 winter (DJF) precipitation variability over the North Atlantic and adjacent Europe.

1.2.2 Solar irradiance

Variations in the solar irradiance influence Earth's climate, albeit in a complex way through, for instance, water vapor feedbacks (Rind et al. 1999; Lean 2000). The Sun's irradiance depends on solar activity and, therefore, on the sunspot number. During the Spörer Minimum (ca. 1420–1570), the Maunder Minimum (ca. 1645–1715), and the Dalton Minimum (ca. 1790–1820) the solar activity was low (e.g. Luterbacher et al. 2001; Stendel et al. 2006). The meridional temperature gradient in the upper troposphere is thereby weakened, due to a cooling of the tropic and midlatitude troposphere related to reduced solar irradiance (Shindell et al. 1999, 2001). The Hadley cell, the zonal winds in the lower stratosphere, and the polar vortex have been found to be weakened during such periods, potentially favoring a negative state of the NAO (e.g. Rind et al. 1999; Shindell et al. 1999, 2001; Wanner et al. 2000, 2001; Hurrell et al. 2004; Luterbacher et al. 2004; Raible et al. 2005; Yoshimori et al. 2005). Recently, the importance of variations in the solar irradiance for European climate has been much debated (e.g. Bengtsson et al. 2006; Stendel et al. 2006). However, various publications emphasize its importance for the downward trend of the temperature during the Maunder Minimum (e.g. Mann et al. 1998, 1999; Lean 2000; Shindell et al. 2001; Waple et al. 2002; Yoshimori et al. 2005), making it also a potential cause for trends in winter precipitation. Therefore, we include this forcing in our study.

2. Data and methods

The seasonal precipitation reconstruction by Pauling et al. (2006) back to AD 1500 covers the European land areas (30° W–40° E; 30°–71° N) resolved on a $0.5° \times 0.5°$ grid. As predictors, long instrumental precipitation series, precipitation indices based on documentary evidence, and precipitation sensitive natural proxies (tree-ring chronologies, ice cores, corals and a speleothem) were used (see Pauling et al. 2006).

The monthly precipitation reanalysis by Mitchell and Jones (2005) served as predictand in the statistical validation. For the 20th century, the reconstruction was updated with this gridded interpolation.

Transfer functions derived from Principal Component Regression (PCR) between the predictors and predictands during the calibration period were applied on the predictors over the period 1500–1900 to obtain the reconstruction. For details on the methodology, we refer to Pauling et al. (2006) and earlier studies using the same methodology (e.g. Luterbacher et al. 2002, 2004; Casty et al. 2005). As every statistical reconstruction, this dataset is associated with uncertainties related to the predictors themselves, the statistical model applied (e.g. unresolved variance within the calibration period), and the dataset used for calibration (Pauling et al. 2006).

2.1 NAO Index and solar forcing

In this study, the NAOI reconstruction by Luterbacher et al. (2002) covering the last 500 years is used. It is based on a combination of long instrumental series, as well as documentary and natural proxy information. This reconstruction is completely independent from Pauling et al. (2006), i.e. they have no predictors in common. Luterbacher et al. (2002) define their NAOI as the standardized difference between the sea-level pressure (SLP) averaged over four grid points on a $5^\circ \times 5^\circ$ grid over the Azores and Iceland, respectively. The reconstruction was replaced by instrumental data from 1901 onwards (Trenberth and Paolino 1980; updated).

As a representation of past solar irradiance, the annually resolved dataset by Lean (2000) is used. This reconstruction covers the period of 1610–2000 and is a combination of direct sunspot observations and information derived from $\delta^{10}Be$ isotopes in ice-cores.

2.2 Selection of the areas and the season of interest

The skill of the Pauling et al. (2006) reconstruction has been estimated using the Reduction of Error (RE) method by Cook et al. (1994). RE

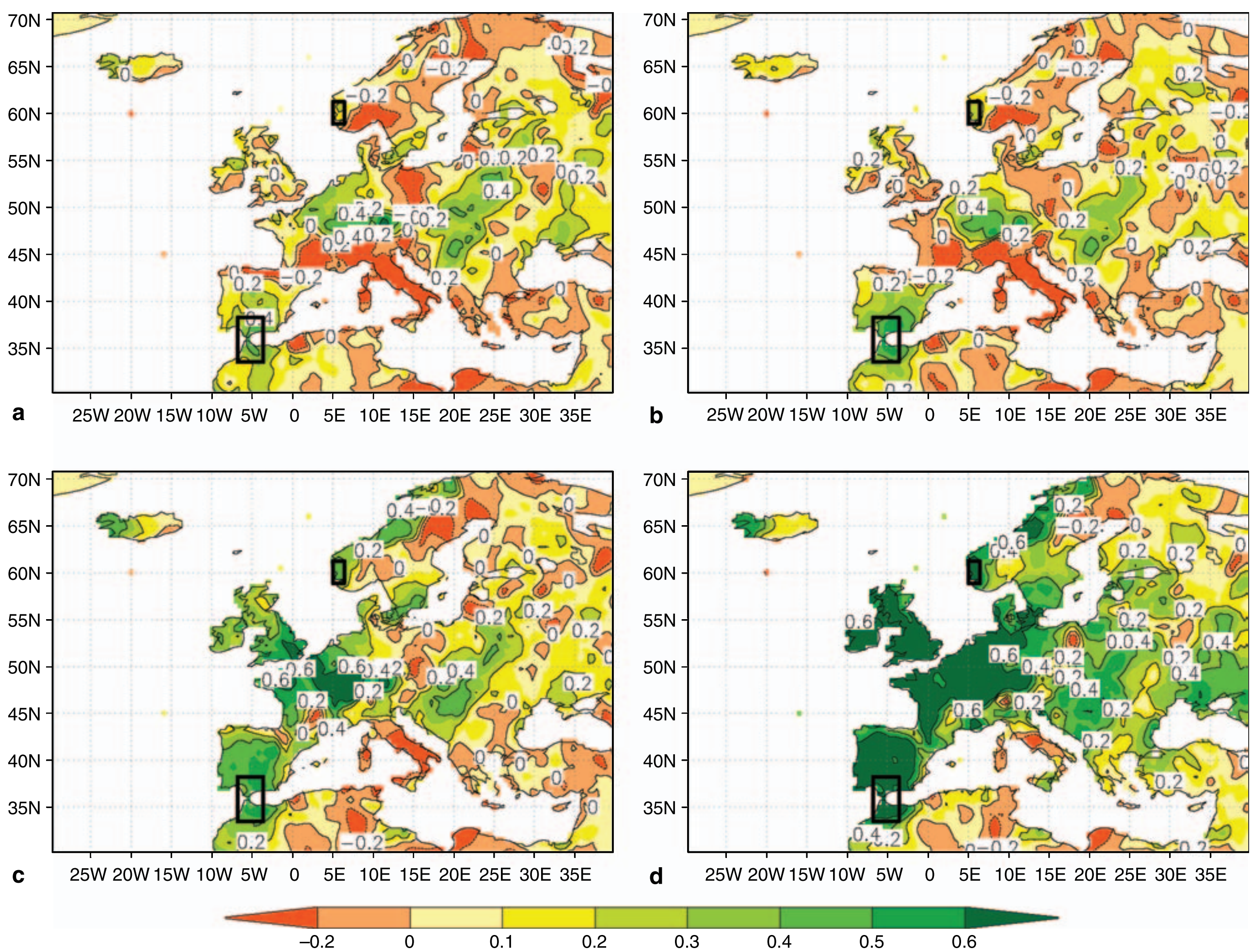

Fig. 1. Spatial distribution of the RE values for the European winter precipitation reconstruction by Pauling et al. (2006) averaged over the (**a**) 16th, (**b**) 17th, (**c**) 18th, and (**d**) 19th century. The black rectangles represent the analysed regions (southwestern Norway 5°–6.5° E; 59°–61° N and southern Spain/northern Morocco 3°–6° W; 34.5°–38° N). Modified after Pauling et al. (2006)

ranges from $-\infty$ to $+1$ with 0 being the skill of climatology. Increasingly positive RE values represent increasing reconstructive skill. $\mathrm{RE}=1$ means perfect reconstruction, i.e. no difference between predictand and reconstruction during the verification period. A reconstruction with $\mathrm{RE}=-1$ is equivalent to a random guess (Cook et al. 1994).

Figure 1 presents the spatial RE values for the European winter precipitation reconstruction by Pauling et al. (2006) averaged over each century. An increase in the reconstruction skill can be observed between 1500 and 1900, coinciding with an increase in the number of available predictors, their quality as well as the predictor network density (Luterbacher et al. 2002, 2006; Jacobeit et al. 2003a; Pauling et al. 2006; Küttel et al. 2007). During the 16th century, only a few regions show RE values over 0.5 (Fig. 1a). Until the 18th century, low reconstruction skill is visible for large parts of Europe, making it difficult to detect regions with positive RE values over the entire 500 years.

In this study, two regions were selected for the 500-year winter precipitation trend analyses (Fig. 1; black rectangles): southwestern Norway (12 grid points) and southern Spain/northern Morocco (36 grid points). Both regions show RE values well above 0 during the entire period. Moreover, southwestern Norway as well as southern Spain/northern Morocco are known to be sensitive to the NAO (e.g. Hurrell and van Loon 1997; González-Rouco et al. 2000; Wanner et al. 2001; Trigo et al. 2004), though with decreasing influence towards the eastern and southern parts of Morocco (Knippertz et al. 2003). Hence, they are interesting from a dynamical point of view.

Only the winter (here defined as the sum of DJF precipitation) is considered, due to several reasons: first, the most reliable reconstruction is obtained during this season because of the good predictor availability (Pauling et al. 2006). More precipitation signals of remote regions simplify the skilful reconstruction in areas with no predictor information, which is the case in southwestern Norway. Thus, the signals are indirectly communicated through teleconnections over the whole reconstructed area (Pauling et al. 2006). Second, the NAO is most pronounced in winter (Dai et al. 1997; Hurrell and van Loon 1997; New et al. 2001; Wanner et al. 2001; Raible et al. 2005). Third, the dynamical relationship between pressure and precipitation is best developed in this season, as precipitation tends to be rather induced by advective than convective processes (González-Rouco et al. 2000; New et al. 2000; Wanner et al. 2000; Frei and Schär 2001; Giorgi 2002; Xoplaki et al. 2004; Zveryaev 2004; Efthymiadis et al. 2006; Casty et al. 2007; Pauling and Paeth 2007).

2.3 Spatial representativeness

To get single precipitation series for each region, the gridded precipitation reconstruction was area-averaged over the 12 (36) grid points of southwestern Norway (southern Spain/northern Morocco). We have calculated the spatial representativeness as used, for instance, by Dai et al. (1997), New et al. (2000), Rapp (2000), and

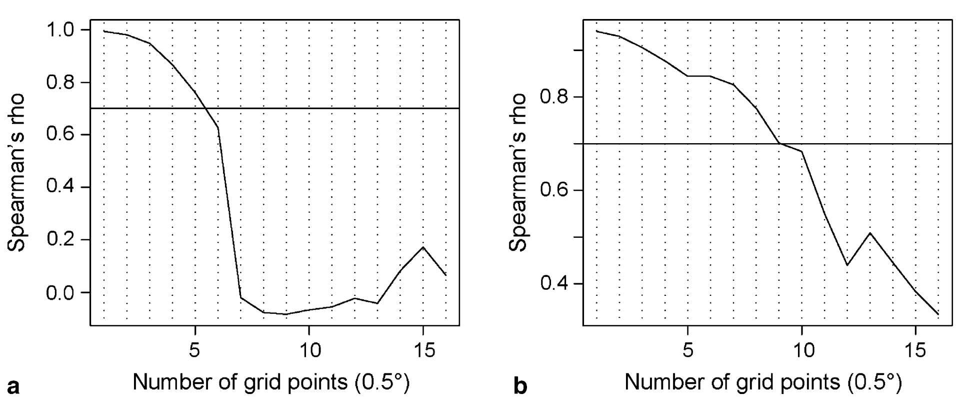

Fig. 2. Spatial representativeness d_R for winter precipitation of (**a**) southwestern Norway and (**b**) southern Spain/northern Morocco during the 20th century. The curve displays the averaged Spearman correlation coefficients as a function of distance. Values below the horizontal line explain less than 50% of the variance. 0.5° of longitude are equivalent to 28 (45) km in southwestern Norway (southern Spain/northern Morocco)

Groisman et al. (2005) to determine the area for which the two regions may provide significant information about precipitation trends. For this purpose, the Spearman rank-correlation coefficients (*rho*) between the precipitation series of the selected regions and the surrounding grid points were calculated (Fig. 2). Rapp (2000) argues that a significant representativeness is considered at values of $r = 0.7$ (for $N = 100$), thus 50% of the common variance being explained. The corresponding distance is called spatial representativeness d_R.

2.4 Trend analyses

We define a trend by the mean slope of the changes in precipitation during a particular time period. The regression coefficient β was obtained from a linear regression using a least squares fit.

Since the starting point of a trend is usually unknown (Dose and Menzel 2004; Percival and Rothrock 2005), the determination of the significance of trends is not straightforward. Various techniques to detect trends and to determine their significance have been used (e.g. Trömel and Schönwiese 2007). However, Moberg et al. (2006) note that there is no "best" technique to address this issue.

We calculate the significance of the trends using the trend test following Mann (1945) and Kendall (1970). According to Rapp (2000), the equation reads as follows:

$$Q = \frac{\sum_{i=1}^{N-1}\sum_{j=i+1}^{N} sgn(y_j - y_i)}{\sqrt{\frac{1}{18}\left[N(N-1)(2N+5) - \sum_{l} b_l(b_l-1)(2b_l+5)\right]}}$$

N is the length of the time series; y_i and y_j stand for pairs of data, which are subtracted from each other; the factor $1/18$ is due to the standardization procedure for $N > 10$; b_l indicates the number of identical observational data y, representing the modification of the Mann-formula by Kendall (1970). Finally, Q is the approximately standard normally-distributed (two-tailed) test value.

This test assumes neither normal distribution of the data nor linearity in the trend (Rapp 2000). It can also deal with missing values. The basic idea of the test is to add up the signs of all possible differences, thus to assess the relative increase or decrease of data elements of a time series. $1/2N(N-1)$ indicates the maximal number of pairs of data (y_i; y_j) with $i < j$ (Rapp 2000). The same formula can be used to calculate the maximal number of combinations in trend matrices (see below).

In our analyses, we use for illustration triangular-shaped matrices as presented by Rapp (2000) and Dose and Menzel (2004). The most important advantage of such a trend matrix is that all possible combinations of time windows of a single time series can be displayed in one figure (Figs. 5 and 6). Within these subintervals the trends and their significance are calculated as described above. Therefore, the problem of time-dependency of the trends as mentioned by Hunt and Elliott (2006) does not exist in our approach. Furthermore, most of the cautionary notes like the unidentified time where the trend starts (Percival and Rothrock 2005), or the inability of single trends to tell whether the data contain one or more trend pattern (Hannachi 2007) can be avoided. With trend matrices any short-term trend can be placed in a long-term framework. This is important as the characteristics of the precipitation trends vary depending on the duration of the trend (Hunt and Elliott 2006).

2.5 Running correlations

Running correlations are widely used in climate research (e.g. Gershunov et al. 2001; Brázdil et al. 2002; Casty et al. 2005; Touchan et al. 2005; Raible et al. 2005, 2006; Luterbacher et al. 2006; Brönnimann et al. 2007). They help to investigate the stability of relations between various climatic variables (Pauling et al. 2006). We calculate the Spearman rank-correlation coefficients between the precipitation time series and the climate indices defined in Sect. 2.1, using a 30-year time window as standard period for calculating climatological means (e.g. New et al. 2000; Rapp 2000; Giorgi 2002; Mitchell and Jones 2005; Raible et al. 2006; Santos et al. 2006). A similar analysis, though on a 200-year window and for global mean temperature, was performed by Mann et al. (1998). The significance is tested applying a z-test.

3. Results and discussion

There are important temporal fluctuations in the winter RE time series of the two selected regions (not shown). The RE values are lowest at the beginning of the reconstruction with values around 0.2 in southwestern Norway and 0.4 in southern Spain/northern Morocco. From 1650 onwards the REs steadily rise to values around 0.8 in southwestern Norway and 0.7 in southern Spain/northern Morocco at the end of the 19th century. The findings of the latter region are closely related to the quality of the available documentary based precipitation indices (Pauling et al. 2006). It is important to note that uncertainties may lead to an underestimation in the RE values for various reasons (e.g. McIntyre and McKitrick 2005). Several studies mention that REs should be interpreted cautiously when the time series are autocorrelated, potentially influencing the significance threshold (e.g. Cook et al. 1994; Gershunov et al. 2001; Percival and Rothrock 2005; Moberg et al. 2006). However, no autocorrelations were found in our winter precipitation series (not shown).

The spatial representativeness d_R (Sect. 2.3) obtained from the instrumental data in the 20th century amounts to around 150 km for southwestern Norway (Fig. 2a) and around 400 km for southern Spain/northern Morocco (Fig. 2b). These values are similar to the findings by Dai et al. (1997) and New et al. (2000). With increasing distance the correlation diminishes rapidly. The lower d_R value for southwestern Norway may be explained by the perpendicular extent of the Norwegian mountains relative to the average flow, i.e. orographically induced meteorological processes.

3.1 500 years of winter precipitation trends

The European winter precipitation reconstruction by Pauling et al. (2006) shows for both regions reduced variability between 1500 and 1700 (Fig. 3), most probably due to the lower number of proxy

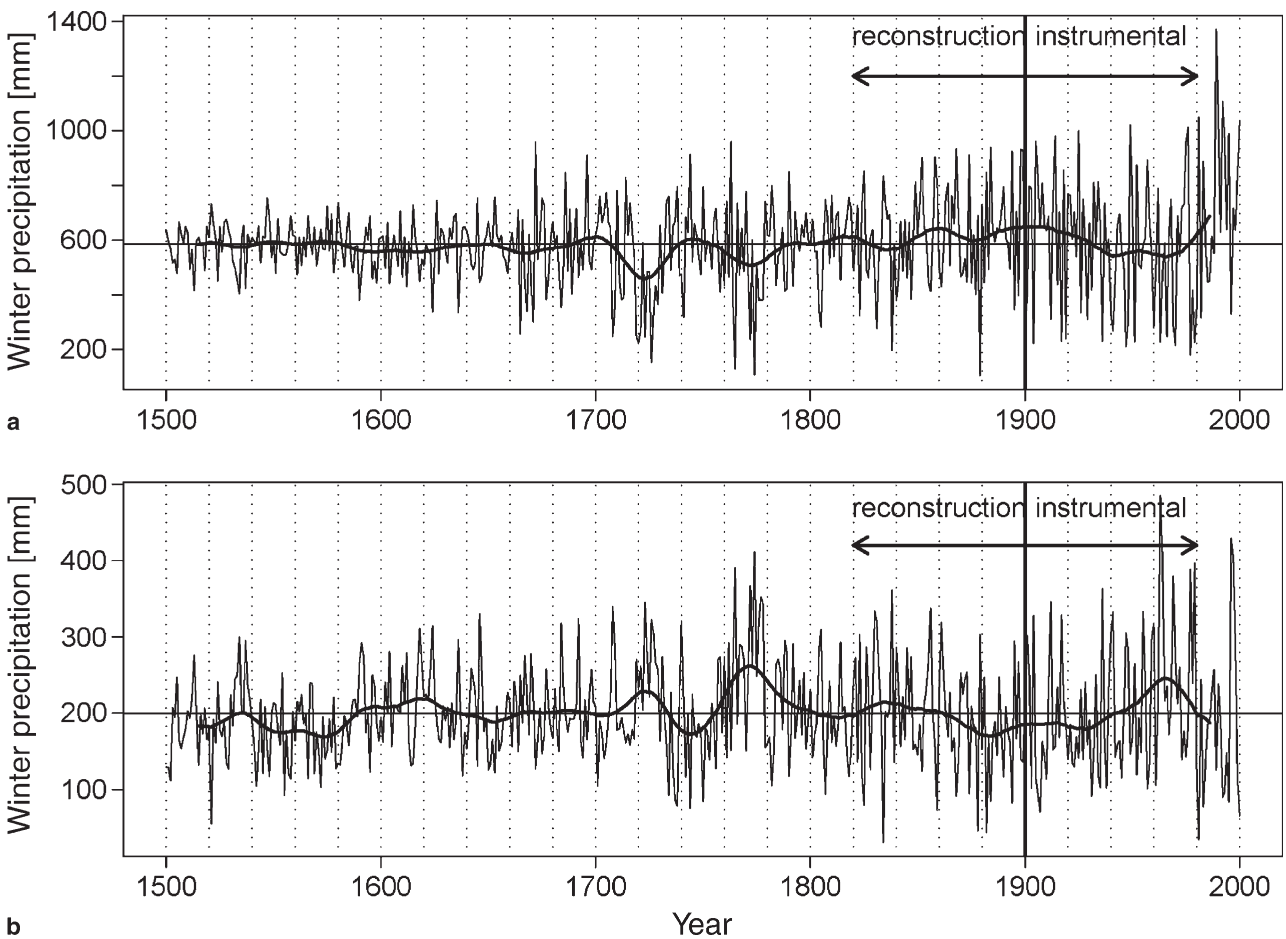

Fig. 3. Reconstructed (1500–1900) and instrumental (1901–2000) winter precipitation time series for (**a**) southwestern Norway and (**b**) southern Spain/northern Morocco, smoothed by a 30-year Gaussian low-pass filter. The horizontal line indicates the 500-year mean

data, their quality, and spatial distribution (e.g. Luterbacher et al. 2006; Pauling et al. 2006; Solomon et al. 2007). This strongly affects analyses of extremes, but to a much smaller extent the means (see Pauling and Paeth 2007), as the predictors were fitted to the mean during the reconstruction process. Additionally, the amplitude of long-term regression coefficients is primarily determined by the mean of the data (e.g. Rapp 2000). Therefore, the reduced variability at the beginning of the reconstruction should not remarkably influence our trend analyses.

Not surprisingly, the 500-year precipitation mean is higher in southwestern Norway (∼590 mm; ∼171 mm standard deviation [SD]; Fig. 3a) than in Spain/northern Morocco (∼200 mm; ∼69 mm SD; Fig. 3b). Both values do not change if the mean is calculated for the 20th century. The standard deviation on the other hand increases in the last century in both regions. As mentioned above, this is primarily due to the increasing reconstruction uncertainty back in time. It is obvious that the last few decades were wetter (drier) than preceding decades in southwestern Norway (southern Spain/northern Morocco), agreeing with the findings of the latest IPCC report (Solomon et al. 2007). The Gaussian low-pass filter underlines the course in mean precipitation over southwestern Norway and particularly over southern Spain/northern Morocco during the 18th century (Fig. 3). Finally, it has to be noted that the variation coefficients are similar in both regions (around 45% during the 20th century; not shown).

We analyse in the following the two precipitation time series on trends. Figure 4 shows the trend curves for both regions for a 30-year moving window. Each regression coefficient is plotted in

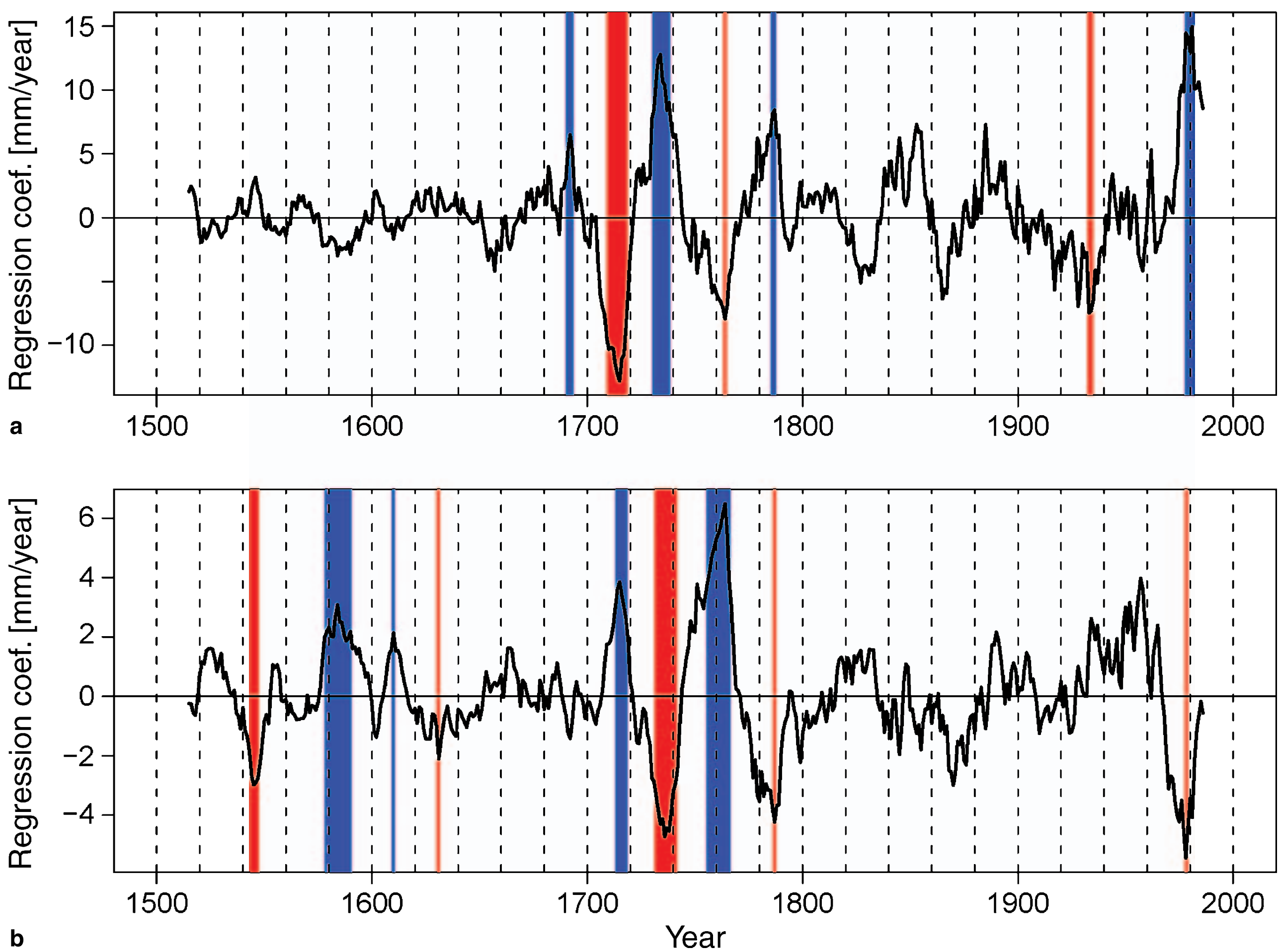

Fig. 4. Winter precipitation trend curves for (**a**) southwestern Norway and (**b**) southern Spain/northern Morocco using a 30-year moving window. Blue (red) columns mark significant positive (negative) trends ($\alpha = 10\%$). Note the different scales. As an example, there was a significant trend to wetter conditions in southwestern Norway during the period 1721–1750 with an increase of around 13 mm/year (given at the year 1735 in the upper panel), i.e. 390 mm more precipitation in 30 years

the middle of the window. Significant trends (at a level of 10%) are overlaid by columns, colored in blue (red) for positive (negative) values. The absolute trend values are larger in southwestern Norway (Fig. 4a) than in southern Spain/northern Morocco (Fig. 4b), but the relative trends are somewhat stronger for southern Spain/northern Morocco (not shown). Note that, based on the Mann-Kendall-trend test (see Rapp 2000), the significance does not depend on the absolute value.

Trend curves based on a moving window are only able to detect the trends of a particular time interval. This disadvantage does not exist in trend matrices applied in this study. With $N = 500$, there are 124,750 possible combinations of subinterval lengths and, thus, of different trends (see Sect. 2.4). It is not suitable to calculate precipitation trends in subintervals shorter than 30 years, since large interannual fluctuations prevent an appropriate interpretation (Rapp 2000; Dose and Menzel 2004). Therefore, the range between $N = 1$ and $N = 30$ is omitted in the trend matrices shown in Figs. 5 and 6.

3.1.1 Southwestern Norway

Figure 5 shows the trend matrix for southwestern Norway with positive trends (wetter conditions) colored in blue and negative trends (drier conditions) in red. Significant trends at the 10%-level are dotted in grey. Beginning years (i) of the subintervals are given in the columns and ending years (j) in the rows. Figure 4a contains the absolute trend values for $N = 30$. In Fig. 5, the 1721–1750 trend discussed earlier is also shown ($i = 1721$; $j = 1750$). The point of intersection lying on the line $N = 30$ returns the same significant positive trend as shown in Fig. 4a.

In general, decadal to interdecadal variability is the dominant feature in Fig. 5. Albeit the trends are mostly nonsignificant, this does not necessarily imply the absence of a trend (Frei and Schär 2001), as indicated by the colored periods without grey dots (Fig. 5). The frequent occurrence of significant longest-term trends is noticeable. This can be explained by the characteristics of the Mann-Kendall-trend test, i.e. a significant trend in large sample sizes is easier

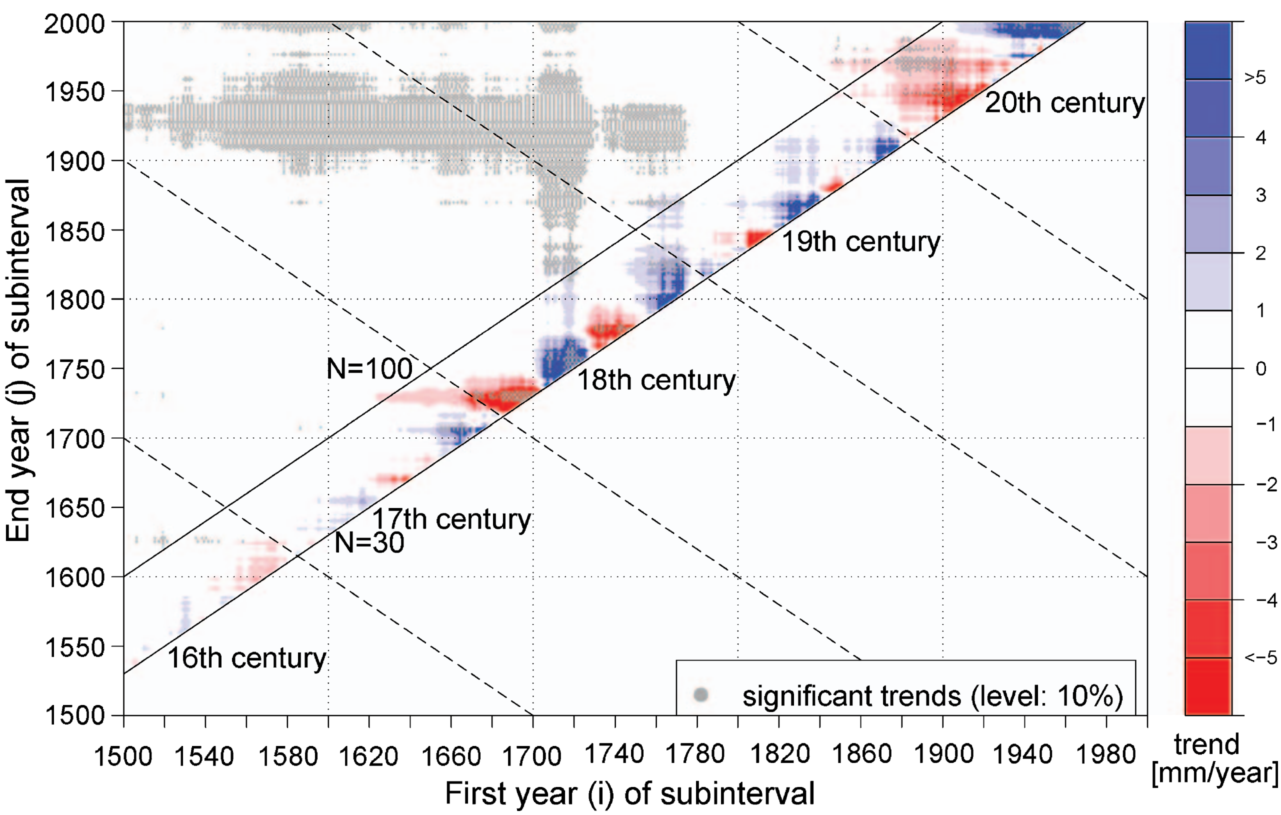

Fig. 5. Trend matrix for winter precipitation over southwestern Norway. Negative (positive) trends are colored in red (blue). Significant regression coefficients ($\alpha = 10\%$) are dotted in grey. The line indicated as $N = 30$ ($N = 100$) represents trends of a 30(100) year long moving window. As an example, the 30-year trend between $i = 1721$ and $j = 1750$ is lying on the line $N = 30$

obtained, since the numerator increases disproportionately compared to the denominator (see equation in Sect. 2.4). However, the trend values are very small (less than 1 mm/year) and therefore not colored.

Between 1950 and 2000, there is a significant trend to more precipitation in southwestern Norway (Fig. 5). Our analyses show for various subintervals positive trend values of ~14 mm/year during this period, the largest coefficients over the entire 500 years (compare with Figs. 3a and 4a). Houghton et al. (2001) and Moberg et al. (2006) found for the 20th century a highly significant increase in the mean precipitation for the European land areas 40°–80° N of +1% per decade. Our larger increase (+3%/decade over the last century; not shown) for southwestern Norway may be due to the smaller size of this region. Other periods with significant positive trends occurred at the end of the 17th century, around the 1740s, and at the end of the 18th century (Fig. 5).

During the 18th and 20th century, significant trends to less precipitation can be found. The time period between 1686 and 1735 contains the most pronounced negative trend of the entire 500 years with a distinct decrease in winter precipitation (up to −250 mm/50 years; Fig. 5). The existence of the negative countertrend around 1960 in the period with the long-term positive trend highlights the problem of making definitive statements about climatic evolution without a clear declaration of a time period (see Hunt and Elliott 2006).

3.1.2 Southern Spain/northern Morocco

Figure 6 shows the trend matrix for southern Spain/northern Morocco. As mentioned earlier, the regression coefficients are smaller in this region than in southwestern Norway.

The significant positive trends in the 16th century are weak, but persist over a long time period. No strong trends were experienced in the 17th and 19th century (compare with Fig. 4b). In the 1760s, the most pronounced positive trends over the last 500 years occurred (+150 mm precipitation in 50 years; Fig. 6). The largest absolute precipitation totals, however, can be found in the 1960s (Fig. 3b). These coincide with a long-

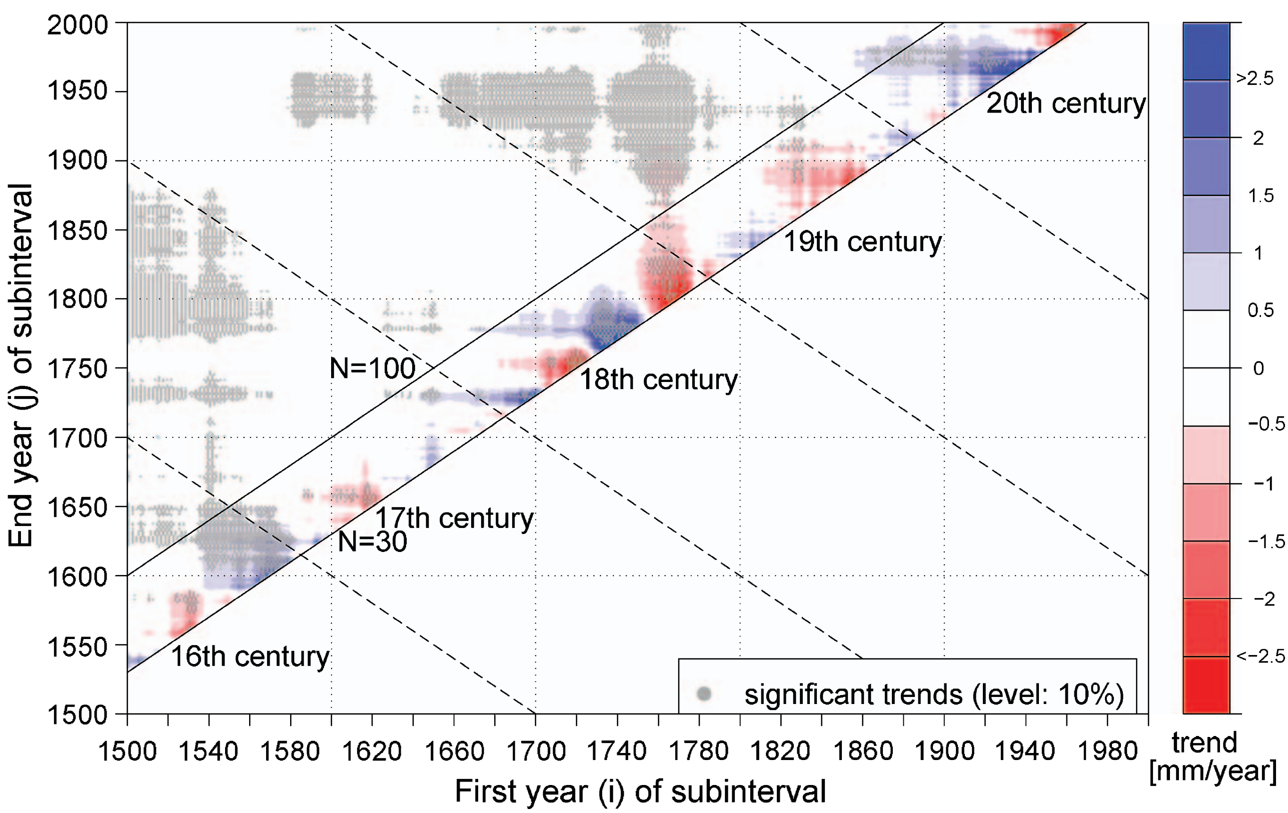

Fig. 6. As in Fig. 5, but for southern Spain/northern Morocco

Table 1. Comparison of the winter precipitation trends of (**a**) southwestern Norway and (**b**) southern Spain/northern Morocco for 50-year periods. Positive (+), strong positive (++), negative (–), and strong negative (– –) trends are distinguished. A positive followed by a negative trend within a 50-year period is marked as "+ –", and vice versa. No sign is drawn if no clear trend occurred. Not significant trends are put in brackets (level at 10%)

	16th century		17th century		18th century		19th century		20th century	
	1501–1550	1551–1600	1601–1650	1651–1700	1701–1750	1751–1800	1801–1850	1851–1900	1901–1950	1951–2000
a		(–)		+	– – ++	–+	(+)		––	++
b	–	+	–		+–	+ + – –		–	++	––

lasting period with significantly more precipitation over southern Spain/northern Morocco around the 1940s.

The positive 1931–1960 trend preceded the recent dryness south of 60° N in Europe (Knippertz et al. 2003; Xoplaki et al. 2004; Solomon et al. 2007). These significant negative trends of the last decades of the 20th century are particularly detectable during various short-term subintervals (e.g. >120 mm/30 years; see Figs. 4b and 6). The large regression coefficients appear to be unprecedented over the last 500 years, what is in agreement with Esper et al. (2007). Beside those of the 20th century, the significant negative trends during the 18th century are remarkable: there was a short period with negative trends in the first half and a larger and stronger one at the end of the century (Fig. 6).

3.1.3 Comparison of the trends of the two regions

To compare the winter precipitation trends of the two regions, we have simplified the trend matrices using signs to distinguish positive and negative trends, i.e. the trends were subjectively summarized over 50-year periods (Table 1). Not surprisingly, the two regions tend to behave in an opposite way. Positive trends in southwestern Norway coincide with negative trends in southern Spain/northern Morocco. The correlation coefficients between the trend curves of the two regions shown in Fig. 4 amount to highly significant values around −0.65 using running correlations, however, being temporally unstable particularly at the end of the 19th century (not shown).

Only southern Spain/northern Morocco shows clear trends in the 16th century (Fig. 6), potentially due to the higher reconstruction skill. During the second half of the 16th century, southwestern Norway was dry, while the winters in southern Spain/northern Morocco were wet (Table 1).

In the first part of the 17th century, only a negative trend in southern Spain/northern Morocco is significant (Table 1). According to Pfister (1999), central Europe was drier than normal (relative to 1901–1960) in the second half of the century. We found a significant positive trend for southwestern Norway during this period (Fig. 5), highlighting the difference of the climate of this region and central Europe. Meanwhile, winter precipitation over the Mediterranean area was normal (Pfister 1999), which is consistent with our results (Table 1).

The end of the Maunder Minimum and the following few decades of the early 18th century contain in both regions the most distinct trend phases over the entire 500 years (Table 1). A negative (positive) trend in southwestern Norway (southern Spain/northern Morocco) followed by an increase (decrease) just as strong is found, particularly in southwestern Norway. Nesje and Dahl (2003) and Nesje et al. (2007) remark the rapid glacier advance between 1710 and 1748 in southern Norway due to increased winter precipitation, which is confirmed by our results. The same evolution from drier (wetter) to wetter (drier) conditions in southwestern Norway (southern Spain/northern Morocco) occurred in the second half of the 18th century, more pronounced in the latter region. Similar results were found by Esper et al. (2007) for Morocco.

Only a few trends are detectable during the 19th century (Table 1). The Dalton Minimum seems not to induce any strong trends. At the end of the century, there was a negative trend in southern Spain/northern Morocco.

According to Table 1, the first half of the 20th century shows more (less) precipitation over southern Spain/northern Morocco (southwestern

Norway), agreeing with Xoplaki et al. (2004). Rodríguez-Puebla et al. (2001) found for the Iberian Peninsula a positive trend around the mid-20th century. Further, Nesje et al. (2007) show a glacier retreat starting in the 1930s in Scandinavia, also agreeing with our results. The second half of the 20th century even contains the strongest negative (positive) regression coefficients over the entire 500-year period in southern Spain/northern Morocco (southwestern Norway). This coincides with the largest glacier advance in southern Norway during the 20th, possibly since the early 18th century (Hurrell 1995; Nesje and Dahl 2003; Nesje et al. 2007).

3.2 Dynamic background

It is extremely difficult to establish causality for many climatic phenomena, mainly for precipitation, because interactions of the various climate factors influence their occurrence (Osborn et al. 2000; Wanner et al. 2000; Casty et al. 2007; Pauling and Paeth 2007). Such analyses are only possible with a profound knowledge of the statistical connections between precipitation and the NAOI or solar activity. However, a significant precipitation trend usually does not occur by chance (e.g. Hunt and Elliott 2006). Therefore, we assess some possible causes behind the detected significant winter precipitation trends. For this purpose, we calculated running correlations using the Spearman rank-correlation method with 30-year time windows between the NAOI/solar forcing (Sect. 2.1) and the reconstructed precipitation series of the two regions. The results were mostly independent of the length of the time window applied (not shown).

3.2.1 Southwestern Norway

Figure 7 presents the 30-year running correlations between the precipitation time series for southwestern Norway and the NAOI and solar irradiance, respectively. Blue (red) columns represent the significant positive (negative) trends in the precipitation series. Obviously, the correlation between the NAOI and precipitation is most of the time positive. From 1780 onwards the correlation is mostly significant (at the 5% significance level). This is in agreement with other studies (Nesje et al. 2001, 2007; Nesje and Dahl 2003). The correlation with the solar irradiance is less significant and more unstable, the coefficients often change sign.

Five of the seven significant precipitation trends in the reconstruction of southwestern Norway coincide with periods with a significant correlation between the NAOI and precipitation time series (Fig. 7). The correlation coefficients are always positive, thus there is usually more (less) precipitation when the NAO is in its positive (negative) phase. Three trends match with a significant correlation between the solar irradiance and precipitation. One of them is only related to the solar forcing. This trend occurred at the end of the Maunder Minimum, i.e. the solar irradiance was increasing. The trends at the end of

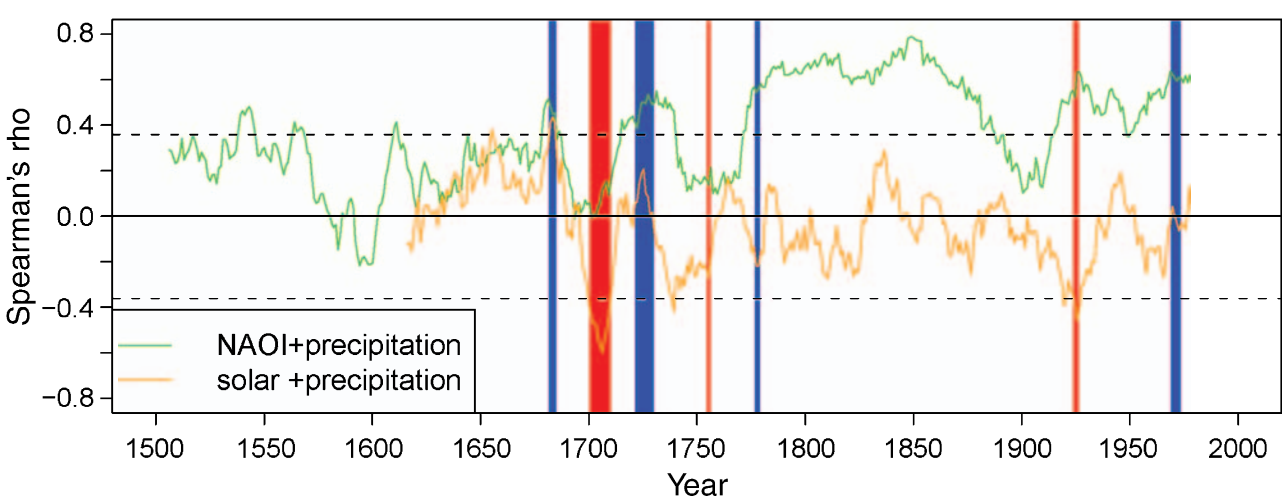

Fig. 7. Running correlations between the winter precipitation of southwestern Norway and the NAOI (green) and solar irradiance (orange) using a 30-year moving window, respectively. Values above/below the dashed line are significant (level at 5%). Blue (red) columns mark significant positive (negative) trends ($\alpha = 10\%$)

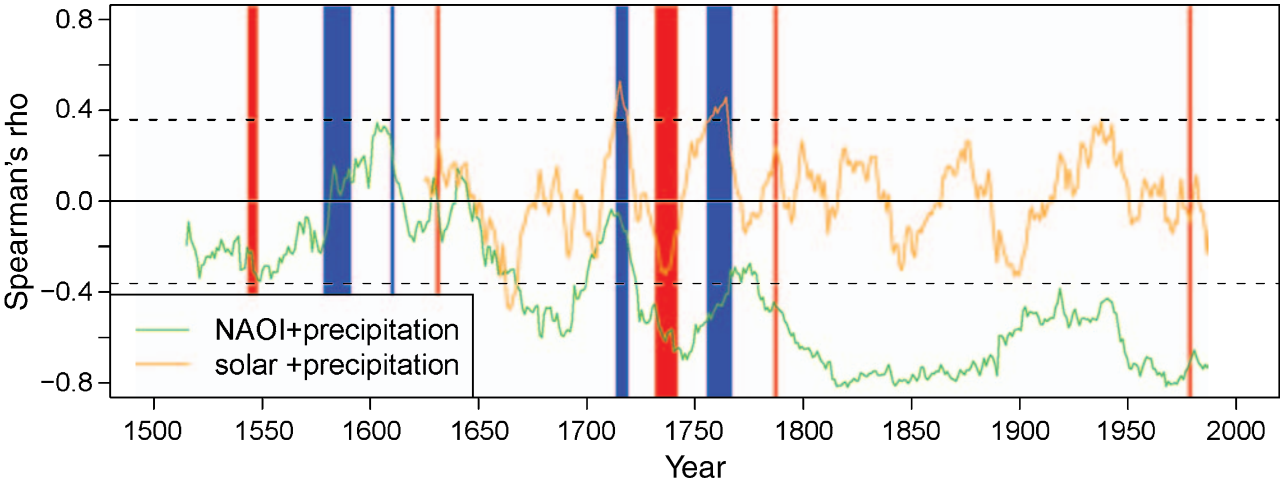

Fig. 8. As in Fig. 7, but for southern Spain/northern Morocco

the 17th century and around the 1940s might be related to the NAO and solar variability at the same time. One significant negative trend in the 1760s remains unexplained.

3.2.2 Southern Spain/northern Morocco

In Fig. 8, the running correlations of southern Spanish/northern Moroccan winter precipitation with the NAOI and solar irradiance are presented, respectively. The NAOI is mostly negatively correlated with precipitation. From 1720 until the end of the 20th century, the correlation is significant, though with a small interruption around 1770 (Fig. 8). The correlation between the precipitation of southern Spain/northern Morocco and solar irradiance is again highly instationary and mostly nonsignificant.

In the 19th century, the correlation coefficients between the NAOI and precipitation are highest (Fig. 8): Over 60% of the precipitation variability can be shared by the NAO alone. Pauling et al. (2006) found similar results using the first Principal Component (PC) of winter SLP reconstruction of Luterbacher et al. (2002). Nevertheless, no significant trends occurred during the 19th century (see Jacobeit et al. 2003a for possible reasons). Four trends coincide with significant NAOI-precipitation-correlations, always with a negative coefficient. More (less) precipitation usually falls during NAO negative (positive) states. The two only periods with significant running correlations between precipitation and solar irradiance over the entire 500 years coincide with significant negative trends, both of them during the five decades following the end of the Maunder Minimum. In the 18th century, the solar irradiance and the NAO appear to be jointly responsible for the significant trends. The significant trends in the 16th and 17th century are not explainable by the two forcings. This may be rather explained by lower SLP and precipitation reconstruction skill than a missing climatic signal (Pauling et al. 2006).

3.2.3 Comparison of the two regions

As a comparison of the two regions, Table 2 shows the significant winter precipitation trends along with the significant correlations between winter precipitation and the NAOI and solar irradiance, respectively. Particularly the 18th and the second half of the 20th century were dominated by internal oscillations in both regions. Evidently, the NAO seems to be the most important reason behind the significant trends on 30-year intervals. Moreover, Table 2 clearly shows the oppositely evolving precipitation pattern over southwestern Norway and southern Spain/northern Morocco.

The significant positive trend in southwestern Norway during the second half of the 17th century coincides with significant positive correlations between precipitation and the NAOI as well as solar irradiance (Table 2). This period includes the early Maunder Minimum. However, no significant correlations were found for southern Spain/northern Morocco during that period.

Table 2. Comparison of the NAO and solar irradiance with the winter precipitation trends of (**a**) southwestern Norway and (**b**) southern Spain/northern Morocco for 50-year periods. Significant positive (+) and negative (–) trends are distinguished using a 30-year moving window ($\alpha = 10\%$). Positive (negative) correlations between the precipitation and the NAOI or solar irradiance are marked as dots [•] (circles [○]). Signs in brackets indicate weak significant correlations (level at 5%). The "?" represents no agreement between the trends and the NAOI or solar irradiance

		16th century		17th century		18th century		19th century		20th century	
		1501–1550	1551–1600	1601–1650	1651–1700	1701–1750	1751–1800	1801–1850	1851–1900	1901–1950	1951–2000
a	trend				+	−+	−+			−	+
	NAO				•	•	? •			•	•
	solar				•	○	?			○	
b	trend	−	+	+−		+−	+−				−
	NAO	(○)	?	? ?		○	○ ○				○
	solar		?	? ?		•	•				

Various recent studies found a dominant negative NAO during the period of the Late Maunder Minimum (ca. 1675–1715; Luterbacher et al. 2001; Xoplaki et al. 2001; Jacobeit et al. 2003b; Raible et al. 2006). Consequently, less (more) precipitation fell over southwestern Norway (southern Spain/northern Morocco) in the early 18th century. However, only the correlations between precipitation and solar irradiance are found to be significant in both regions during this time (Table 2). This highlights the problem of overlapping internal oscillations and external forcings as presented in Figs. 7 and 8. Patterns like the NAO may be modulated by volcanic eruptions, solar irradiance, or greenhouse gases (Robock 2000; Luterbacher et al. 2001; Hegerl et al. 2003; Stott 2003; Xoplaki et al. 2004; Brönnimann et al. 2007; Fischer et al. 2007), and/or internal variability can mask an external signal (Yoshimori et al. 2005; Pauling and Paeth 2007).

After the Maunder Minimum, the solar irradiance increased again, accompanied by a positive state of the NAO (Shindell et al. 2001; Luterbacher et al. 2004; Stendel et al. 2006) and, therefore, more (less) precipitation in southwestern Norway (southern Spain/northern Morocco; Table 2). As mentioned in Sect. 3.1.3, the same change from drier (wetter) to wetter (drier) conditions in southwestern Norway (southern Spain/northern Morocco) occurred in the second half of the 18th century (Table 2). Over southern Spain/northern Morocco, the significant positive trend can clearly be related to the solar irradiance and slightly to the NAO (Fig. 8). The causes of the significant negative trend over southwestern Norway remain unclear. At the end of the 18th century, the NAOI-precipitation-correlation and, thus, the precipitation trend is found to be negative (positive) in southern Spain/northern Morocco (southwestern Norway). This agrees with findings by other studies (Wanner et al. 2000; Luterbacher et al. 2002; Casty et al. 2005).

In the middle of the 20th century, the correlation of the NAOI with the precipitation in southwestern Norway is positive. As the NAO was in its negative state (Hurrell 1995; Hurrell and van Loon 1997), this could explain the negative precipitation trend there to some extent. The solar irradiance seems to be jointly responsible (Table 2), although probably not in a causal way. During the last decades of the 20th century, a positive NAO prevailed (Hurrell 1995; Hurrell and van Loon 1997; Wanner et al. 2001; Hurrell et al. 2004; Lohmann et al. 2005; Percival and Rothrock 2005; Raible et al. 2005, 2006; Stendel et al. 2006; Casty et al. 2007). The recent negative trend of the Mediterranean winter precipitation as well as the positive trend over southwestern Norway can at least partly be explained by this NAO state (Giorgi 2002; Trigo et al. 2004; Xoplaki et al. 2004; Luterbacher et al. 2006, 2007). The latter trend is thermodynamically consistent with higher temperatures, as there must be an increase in precipitation to balance the enhanced evaporation caused by the risen temperatures (Dai et al. 1997; New et al. 2001; Giorgi 2002; Esper et al. 2007; Solomon et al. 2007). According to Hurrell (1995), the positive NAO during the last decades of the 20th century contributed significantly to the recent wintertime warmth across Europe.

3.2.4 Not analysed potential influences on significant trends

It is obviously not possible to attribute causes to all significant trends. This may be due to the instationarities in the relationship between precipitation and the NAO or solar forcing (see Figs. 7 and 8), as was also found by Jacobeit et al. (2003a), Trigo et al. (2004), Casty et al. (2005, 2007), Lohmann et al. (2005), Touchan et al. (2005), Pauling et al. (2006), Raible et al. (2006) and Brönnimann et al. (2007). The reasons for this are numerous: Lohmann et al. (2005) point to the last 50 years which may not be representative in the long-term context due to shifts in teleconnection patterns, and may not be suitable as a calibration period for reconstructions (see Wanner et al. 2001). Hunt and Elliott (2006) revealed a near-Gaussian distribution of the trends, i.e. stochastic processes dominate climatic variability, thus also the trends (see Frei and Schär 2001; Gershunov et al. 2001; Bengtsson et al. 2006). Additionally to the above mentioned uncertainties in the precipitation reconstruction involving the underestimation of low-frequency variability, uncertainties in the NAOI and solar reconstruction used in this study may lead to instationarities.

Further, there are parts of the variance which may be explained by aspects not considered in our analyses. For example, it is possible that naturally occurring modes of atmospheric variability (e.g. the NAO) may be anthropogenically influenced or by unknown factors (Shindell et al. 1999, 2001; Waple et al. 2002; Hurrell et al. 2004). Zhang et al. (2007) estimate that anthropogenic forcing contributed significantly to observed increases in annual precipitation in the Northern Hemisphere mid-latitudes and to drying in the Northern Hemisphere subtropics.

Furthermore, the ENSO could also influence the precipitation distribution in the North Atlantic-European region. Cassou and Terray (2001) or Lohmann et al. (2005) found a negative NAO state during La Niña conditions. This result is not confirmed by Brönnimann et al. (2007), who found a consistent and statistically significant ENSO signal for late winter and spring climate after removing years following tropical volcanic eruptions. They suggest a link between El Niño events and negative NAO modes. Our analyses (with different time-lags) revealed no clear influence of the El Niño/Southern Oscillation on the precipitation trends of both regions (not shown). This is in accordance with other studies (e.g. Rodríguez-Puebla et al. 2001; Zveryaev 2004).

Finally, Bengtsson et al. (2006) found from temperature model simulations using fixed pre-industrial forcing that climate variability in Europe during the pre-industrial period is fundamentally a consequence of internal fluctuations of the climate system. Only a major external forcing could have an important and direct influence on the European climate (Bengtsson et al. 2006).

4. Conclusions and outlook

The dataset from Pauling et al. (2006) is the first gridded 500 year long seasonally resolved precipitation reconstruction over European land areas. It provides a useful basis on which past precipitation variability can be analysed within a long-term context. Analyses of trends allow insights into the characteristics of climate variations on different spatio-temporal scales (Frei and Schär 2001), being of major interest particularly for precipitation as a key climatic element in ecosystems and societies (e.g. Schmidli et al. 2002; Xoplaki et al. 2004; Pauling et al. 2006).

We have performed trend analyses, based on the 500-year winter precipitation reconstruction. Trend matrices have proven to be a suitable method to detect trends of a time series on all possible subintervals and making them visually accessible. The trend detection was investigated for two regions with a high reconstruction skill over the entire half millennium, namely southwestern Norway and southern Spain/northern Morocco. Additionally, the selection was dynamically motivated as these two regions are oppositely influenced by the NAO (e.g. Hurrell and van Loon 1997). This signal can also be shown in our trend analyses. Differences in the timing and strength of the trends of these regions were found, with e.g. southwestern Norway (southern Spain/northern Morocco) showing more significant positive (negative) winter precipitation trends. In both regions, the most pronounced trends occurred during the 18th century with negative trends followed by positive countertrends. The positive (negative) regression coefficient over southwestern Norway (southern Spain/northern Morocco) during the last few decades of the 20th century is unprecedented in the context of the last 500 years. Furthermore, the positive trend in northern Europe was linked with the largest glacier advance in southern Norway during the 20th century (Nesje and Dahl 2003; Nesje et al. 2007) and an unprecedented warmth for at least the past 500 years (Luterbacher et al. 2004, 2007).

In order to find possible causes of the significant trends, we have used running correlations between the precipitation series of the two regions and the NAOI and solar irradiance, respectively. Major instationarities were revealed for both regions. Along with the uncertainties in the precipitation and NAOI data, especially in the pre-1800 period, every reconstruction has to be interpreted with caution, and the exact numerical trend values should be treated with care. However, the NAO appears to be the dominant reason for significant trends in the two selected regions. Although external forcings could be masked by internal oscillations (e.g. Yoshimori et al. 2005) or may be irrelevant for pre-industrial European climate variability (Bengtsson et al. 2006), most of the significant trends during the Maunder Minimum coincide with periods of significant running correlations between precipitation and solar irradiance. Other causes behind the significant trends not considered in our analyses could be anthropogenically induced effects, such as the greenhouse gas forcing particularly during the 20th century (e.g. Zhang et al. 2007), or internal large-scale oscillations like the ENSO.

Future studies may include other seasons as well as other regions. Further, it would be interesting to compare the trends found in our study with results obtained by model simulations. Finally, another method to detect trends could be used (e.g. Trömel and Schönwiese 2007).

Acknowledgements

We wish to thank Jürg Luterbacher (University of Bern) for comments and fruitful discussions. This work uses results from the EU project SO&P (Simulations, Observations and Paleoclimatic data: climate variability over the last 500 years) and was finalised in the Swiss NCCR Climate project PALVAREX (Paleoclimate Variability and Extreme Events). Marcel Küttel was supported by the Swiss National Science Foundation (NCCR Climate), Andreas Pauling by the EU project SO&P. This publication has been funded by the Foundation Marchese Francesco Medici del Vascello.

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