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Interannual variability of winter precipitation in the European Alps: relations with the North Atlantic Oscillation

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Abstract

The European Alps rely on winter precipitation for various needs in terms of hydropower and other water uses. Major European rivers originate from the Alps and rely on winter precipitation and the consequent spring snow melt for their summer base flows. Understanding the fluctuations in winter rainfall in this region is crucially important to the study of changes in hydrologic regime in streams and rivers, as well as to the management of their water resources. Despite the recognized relevance of winter precipitation to the water resources of the Alps and surrounding regions, the magnitude and mechanistic explanation of interannual precipitation variability in the Alpine region remain unclear and poorly investigated. Here we use gridded precipitation data from the CRU TS 1.2 to study the interannual variability of winter alpine precipitation. We found that the Alps are the region with the highest interannual variability in winter precipitation in Europe. This variability cannot be completely explained by large scale climate patterns such as the AO, NAO or the EA-WR, even though regions below and above the Alps demonstrate connections with these patterns. Significant trends were detected only in small areas within this region, and were of opposite sign between the eastern and western part of the Alps.

1 Introduction

European Alps (43° N÷49° N, 4° E÷19° E) are characterized by a complex climatology, due to the orography, the geographical location, and the interactions with the weather systems which move eastwards from the Atlantic Ocean. The mountain chain stands at the crossroad of many different climatic systems (polar, atlantic, saharan, mediterranean and continental) and, because of its great altitudinal range, it exhibits a variety of climatic regimes similar to that observed across widely separated latitudinal areas. The abundant water resources of the Alpine region are contributed both by rainfall and by snow, with snow being a strategic seasonal storage of water that becomes available in the warm season, when snowmelt provides water both for agriculture and industrial uses (i.e. mainly, for irrigation and hydropower). Because of the vulnerability of the Alps to climate change (Beniston et al., 1997) and their contribution to aquifer recharge and to the base flow of the main European rivers, it is important to understand the patterns and drivers of seasonal and interannual changes in precipitations, especially during the winter season, when most of the water storage accumulates in the snow pack. This study investigates fluctuations and trends in the interannual variability of precipitation and seeks for a relationship between monthly/seasonal precipitation and large-scale patterns of atmospheric circulation. To this end, we assess the strength and areas of influence of possible climatic teleconnections and investigate the specific role that the Alps play in the European climatology. Although in recent years a number of authors have investigated the patterns of interannual variability of precipitations in the Alps (Quadrelli et al., 2001; Beniston and Jungo, 2002; Schmidli et al., 2002), these patterns and their relation with large-scale climate oscillations have never been assessed using spatially extended climate records, and in some cases only relatively short series were used (e.g., Quadrelli et al., 2001). Indeed, the use of long time series is crucially important to the analysis of climate trends and low-frequency modes of climate variability. The modes of atmospheric circulation considered in this study include the North Atlantic Oscillation (Hurrell, 1995), the Arctic Oscillation (Thompson and Wallace, 1998) and the East Atlantic West Russia (Barnston and Livezey, 1987). These large-scale patterns affect the weather and climate of wide geographical regions in the Northern Hemisphere (Hurrell and Van Loon, 1997; Thompson and Wallace, 2001). The North Atlantic Oscillation (NAO), (Hurrell, 1995), is a large-scale pattern in the winter season atmospheric circulation in the North-Atlantic. The positive phase of the NAO is associated with anomalous high pressure in the subtropics and low pressure in the subarctic, with stronger westerly winds and enhanced flow of moist and warm air through the North Atlantic and Western Europe. Due to the stronger westerlies, (Lamb and Peppler, 1991; Hurrell, 1995) winter precipitation is higher than average in the region between Scandinavia and Iceland and lower than average in Southern Europe.

Winter temperatures are also affected by the NAO with the flow of cold arctic air across western Greenland and the North-Western Atlantic, and the enhanced westerly flow of warm air over Europe. These conditions lead to relatively warm and rainy winters in Northern Europe (Hurrell, 1995). Opposite conditions occur during the low phase of the NAO. These patterns have been related to temperature fluctuations in Eurasia and to the strength of the stratospheric polar vortex, which are associated with the Arctic Oscillation (e.g., Thompson and Wallace, 1998, 2000) and the East Atlantic-West Russia pattern (Barnston and Livezey, 1987).

2 Data

The study of climate patterns in the relatively large and complex region of the European Alps requires the use of complete and long time series of spatially distributed monthly precipitation. To this end, a gridded dataset, constructed interpolating station records on a grid of given spatial resolution, was used. Among the existing databases (Global Air Temperature and Precipitation, NOAA- Center for Climatic Research Department of Geography University of Delaware, 2001; Analysis (Schmidli et al., 2001); HISTALP – ALP – IMP (Auer et al., 2005), we selected the monthly precipitation time series of the gridded database CRU TS 1.2 (Mitchell et al., 2003), developed by the Tyndall Centre for Climate Change Research and the Climate Research Unit (CRU) of the University of West Anglia. This data set provides the longest time series and a good spatial resolution (grid spacing of about 20 km). The dataset includes monthly time series of precipitation, temperature, vapour pressure, diurnal temperature range and cloud cover for all Europe ($34^\circ \div 72^\circ$ N, $11^\circ \div 32^\circ$ E) for the period 1901–2000. The dataset has been constructed with the anomaly approach (New et al., 1999) interpolating station data with a procedure that considers, as parameters, latitude, longitude and elevation. The advantage in using this type of gridded dataset is that we can handle long and uninterrupted time series for all the grid points. However, the interpolation procedure can induce some errors and bias in the gridded data, especially in areas with topographic singularities and low station density. To assess the ability of this dataset to capture the full variability of real data, we used data records from nine meteorological stations and checked the consistency between station records and gridded data in the grid cells. The station data (Table 1) were taken from the archives of Regione Piemonte, Italy, and from the Global Historical Climatology Network of the National Climatic Data Center (National Oceanic and Atmospheric Administration, USA). The Pearson's correlation coefficient between the time series extracted from the gridded dataset and the station records was used as an indicator of the consistency between gridded and station data. Scatterplots (Fig. 1) were also used to assess the possible existence of a bias (i.e. deviation from a 1:1 dependence). While some stations show a strong correlation as well as a good agreement between station and grid data (Fig. 1a), in other cases, despite the strong correlation, bias was observed, as the scatterplot deviated from the 1:1 line (Fig. 1b). In other stations the data roughly followed the 1:1 line, though with a weaker cross-correlation (Fig. 1c). Moreover, the comparison between the annual and seasonal coefficient of variation of precipitation calculated for the same location using station records and gridded data demonstrates that the CRU TS 1.2 captures the natural variability of precipitation (Table 1), so we can conclude that the variance of original data is well reflected in the gridded database. The temporal patterns of the North Atlantic Oscillation (NAO) were quantified through the NAO Index (NAOI) from the Climate Research Unit of the University of West Anglia (Jones et al., 1997); the NAOI time series covers the 1821–2000 period and represents the normalized sea surface pressure difference between Iceland (Reykjavik) and Gibraltar. The Arctic Oscillation and the East Atlantic West Russia (Eurasian Pattern 2, Barnston and Livezey,1987) have been quantified using indices available from the database of the NOAA, Climate Prediction Center. These indices, available for the period 1950–2000, have been calculated by applying the Rotated Principal Component Analysis (RPCA) to the monthly mean standardized 500-mb geopotential height anomalies.

3 Methods

The spatial distribution of the interannual variability of precipitation was first evaluated through the coefficient of variation, calculated for each grid cell using the time series of the winter season (DJFM) precipitation. The Standardized Precipitation Index (SPI, Mc Kee et al.,1993) was also used as a normalized indicator of the climate variability in the study area; in fact, even if it is generally used as a drought index, it is in general a metric for the study of deviations from the mean, including wet anomalies. The use of the SPI has some important advantages: first, it transforms the precipitation records into a standardized time series (i.e. after subtraction of the mean and division by standard deviation); as such, values of SPI in different place can be readily compared using their frequency of occurrence. Second, it can be calculated with different temporal scales of aggregation and related to events at synoptic scale. Once the aggregation periods of i-months have been selected, the precipitation time series for the i-months can be derived as moving summation of the previous i-months. In order to identify the relationship between probability and precipitation, Mc Kee et al. (1993) suggests to fit a gamma distribution to the aggregated precipitation time series. The SPI is then calculated by means of an equiprobability transformation from the fitted Gamma distribution to the standard normal distribution and represents the difference between precipitation values and the mean, divided by the standard deviation for a given averaging period t. The application of this procedure to the data, using the Anderson-Darling test, shows that the gamma distribution fitted the times series quite well: for 85% of the European Alps the fit passed the test with a significance level of 0.05 for the 3-months aggregation time scale. Moreover, the results obtained with this method did not differ from the anomalies calculated without fitting a Gamma distribution. Therefore we follow the procedure as indicated by Mc Kee et al. (1993). We concentrate only on the winter season data, and calculate the SPI just for three time scales (1, 3 and 6 months) ending in the months of January, February and March. In this way, we obtained three different categories of indices (SPI1, SPI3, SPI6). In addition, to investigate rainfall variability throughout the entire winter season, a modified SPI was determined using aggregation intervals starting in December and ending respectively in January, February and March. This choice was motivated by the fact that the NAO, AO and the EA/WR are known for their effect on winter European climate. Thus, we obtain the $SPI_{m,JAN}$ with the precipitation of December and January, the $SPI_{m,FEB}$ with the precipitation from December through February and the $SPI_{m,MAR}$, that includes all the winter's months (DJFM). Temporal trends in the precipitation and Standardized Precipitation Index time series were tested using the Mann-Kendall test with a 5% significance level (e.g., Helsel and Hirsch, 1992), while Spearman's Rank Correlation Test was used to evaluate the association between climate patterns and precipitation variability (e.g., Helsel and Hirsch, 1992). This method was preferred to the Pearson correlation because it does not need any assumption about the frequency distribution of the variables and it does not require a linear relationship between them. The same test was carried out by removing from the times series years with NAO close to "neutral" (i.e. NAOI in the interval (–1,1)) to assess the effect of more extreme NAO phases on the precipitation variability.

4 Results

Figure 2 shows the coefficients of variation of winter season precipitation across Europe. It is found that the Alps are a singular area of Europe, where the coefficients of variation of winter precipitation (particularly in the Eastern Alps) are observed to be much higher than in the rest of the continent. Although mountainous areas are known for the relatively strong variability of their rainfall regimes presumably due to orographic effects on local climate patterns, the relation between the interannual variability of precipitation and elevation remains unclear (Chacón and Fernandez, 1985). As we did not find a relationship between the coefficient of variation and the mean elevation of the grid cell or the mean winter precipitation able to explain the relatively strong interannual variability observed in the Alps, we investigated its relation with large-scale patterns of

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climate variability such as the NAO, AO, and EA/WR, whose influence on the European climate has already been documented in the literature. In this study we used monthly precipitation and SPI time series to relate interannual fluctuations of alpine precipitation to indices representing the strength of these large scale climate patterns. The Spearman's rank correlation coefficients (Fig. 3) calculated between NAOI and winter precipitation (JFM) demonstrate that in the European Alps the dependence between winter precipitation and the NAO is generally weak. In fact, it appears that the Alps are the European region in which the significance (i.e. p-values) of this relation is particularly low (Fig. 3b), with only a slightly higher significance on the Southern slope of the Alps. While the NAO is known to play an important role in determining climate variability in other areas of Europe such as the Iberian Peninsula, Scandinavia, and the British Isles (Hurrell and Van Loon, 1997), no clear NAO signature can be found in the interannual fluctuations of precipitation in the Alps. The effect of the NAO on the European climates has opposite sign in Northern and Southern Europe: the NAOI is positively (negatively) correlated with winter precipitation in northern (southern) part of the continent. The sign change occurs about at the latitude of the Alps (Fig. 3). Similarly, when compared to the rest of Europe, the Alps exhibit the weakest correlation between the NAOI and the Standardized Precipitation Index (SPI1, SPI3, SPI6 and modified SPI, see methods) as shown in Fig. 4 for the case of SPI with modified time scale. To assess the influence of the extreme NAO phases on precipitation events, we calculated the Spearman's rank correlation eliminating from the NAOI and SPI time series the years with NAOI values in the range (–1,1). The spatial distribution of these correlation coefficients (not shown) exhibits patterns similar to those in Fig. 4 but with slightly higher values, demonstrating that the stronger phases of the NAO have a stronger effect on the European climate. The Alpine region confirms to be a transition region, with weak dependence on NAOI. The view emerging from these analyses is that the patterns of the North Atlantic Oscillation are unable to explain the strong interannual variability of precipitation observed in the Alps. The same statistical analyses used for the North Atlantic Oscillation were applied to assess the dependence of

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precipitation and of SPI on two other patterns of atmospheric circulation, namely, the Arctic Oscillation(AO) and the East Atlantic West Russia. Not surprisingly, the AO has a similar impact on the regional precipitation as the NAO. The influence of the EA/WR is more interesting: both with precipitation and SPI the correlation coefficients found for the alpine region are mainly negative and significant (Fig. 5), especially for the northern and eastern slope, even if the association between the time series does not last until the end of the winter season and ends in February. The presence of trend in winter precipitation was finally assessed using the Mann-Kendall Trend Test. We identified two sub-regions: in the Western part of the Alps the trends were positive, while in the Eastern part they were negative. However, these trends were significant only in small areas (Fig. 6). These results contrast with those obtained by other authors (Quadrelli et al., 2001; Schmidli et al., 2002; Haylock and Goodess, 2004) who found more signi-ficatcant trends in Alpine precipitation. These differences can be presumably ascribed to the different length of the rainfall records (Quadrelli et al.,2001, considers only the period 1971–1992) and to the methods used. For example, Schmidli et al. (2002) calculated the trend as a fraction of the deviation from the climatological mean, while Haylock and Goodess (2004) concentrated only on extreme precipitation events. Our results show (Fig. 7) a significant and spatially coherent negative trend for SPI3 and SPI6 over the Eastern European Alps, indicating a decrease in precipitation and an increase in the occurrence of dry anomalies.

5 Discussion and conclusion

In this study the interannual variability of winter alpine precipitation has been investigated using the monthly precipitation time series obtained from the dataset CRU TS 1.2. Despite the bias induced by the spatial interpolation, overall the interpolated gridded monthly data capture quite well the temporal variability of precipitation observed in the station records. The small differences found can be ascribed to the fact that the time series extracted from the gridded data contain reference precipitation val-

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ues for the entire cell, and its variability is different from that of precipitation measured at a point (e.g., D'Odorico and Rodriguez–Iturbe, 2000). The Alpine region exhibits the strongest interannual variability of winter precipitation in Europe. This variability has important implications on winter Alpine tourism, aquifer recharge, water availability for irrigation and hydropower, and the hydrological regime of major European rivers which rely on spring snow melt for spring and summer flows. This strong year-to-year variability in winter-season precipitation cannot be completely explained by the dominant large-scale modes of climate variability in the Northern Hemisphere. The Arctic Oscillation and the North Atlantic Oscillation have only a weak impact on winter Alpine precipitation. Our analyses show that weak dependences are found regardless of the aggregation scale, the use of time lags between climate pattern and precipitation (not shown), or the analyses of more extreme phases of these climate patterns. The East Atlantic-West Russia shows a significant negative correlation with precipitation anomalies but only for the first part of the winter season. Moreover, as the EA/WR index time series is shorter than those of NAOI and begins only in 1950, the results obtained with the two atmospheric pattern are not comparable. The interannual variability of precipitation and SPI could be partly contributed by trends, though significant trends were observed only in gridded data in the Eastern sector of the Alpine region while they were not even detected in the station data. The significant negative trends found in winter precipitation and SPI in the Eastern Alps indicate a decrease in precipitation and an increase in short-term droughts in this part of the region (Figs. 6 and 7). Overall this study shows that the Alps are a rather singular climatic region in Europe, which exhibits precipitation regimes with two major distinctive features: 1) a particularly high interannual variability of winter precipitation, and 2) a particularly weak dependence on the North Atlantic Oscillation and a slightly stronger association with the East Atlantic West Russia for the first part of the winter. The relatively strong variability of winter precipitation in the Alps seems to be endogenous to this region, and to emerge as a result of the complex interactions between weather system and the topography.

Acknowledgements. The NAO Index has been calculated by et Jones et al., 1997 (available at the web page www.cru.uea.ac.uk/cru/data/nao.htm). The AO and EA-WR indices were calculated by the Climate Prediction Centre (available online: ftp://ftp.cpc.ncep.noaa.gov/wd52dg/data/indices/tele_index.nh). The precipitation time series for the meteorological station have been provided by Regione Piemonte (CD source) and by the National Climate Data Center (available online at http://www.ncdc.noaa.gov/oa/climate/ghcn-monthly/index.php?name=precipitation The dataset CRU TS 1.2 is available online, subject to request to the autors, at the web page http://www.cru.uea.ac.uk/~timm/grid/CRU_TS_1_2.html.

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Table 1. Stations used for the comparison with the gridded dataset

Station	Lat (° N)	Lon (° E)	Elevation (m a.s.l.)	Period of record	r^*	CV Station	CV cell
Alessandria (IT)	44.91	8.62	95	1901–1986	0.821	0.889	0.818
Dijon (FR)	47.26	5.08	227	1831–2006	0.997	0.591	0.575
Domodossola (IT)	46.11	8.29	252	1901–1996	0.727	1.055	0.965
Grenoble (FR)	45.17	5.72	212	1845–1988	0.883	0.628	0.596
Klagenfurt (AU)	46.65	14.32	459	1813–2006	0.971	0.751	0.624
Saentis (CH)	47.25	9.35	2500	1883–2006	0.920	0.666	0.695
Sonnblick (AU)	47.05	12.95	3109	1890–2006	0.809	0.488	0.536
Trento (IT)	46.07	11.12	199	1861–1976	0.964	0.945	1.267
Zürich (CH)	47.38	8.56	569	1836–2006	0.973	0.587	0.609

* Coefficient of correlation between the station and dataset time series.

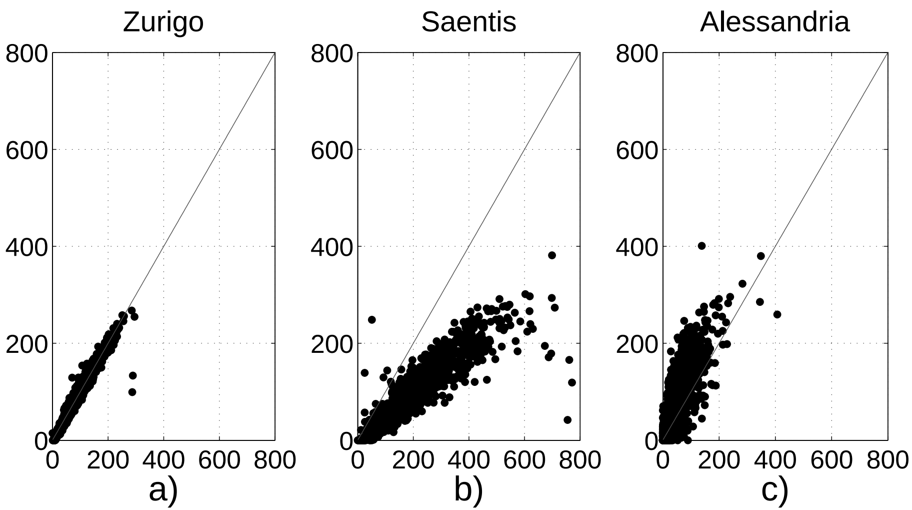

Fig. 1. Representation of the agreement between monthly values of precipitation from the gridded and station data.

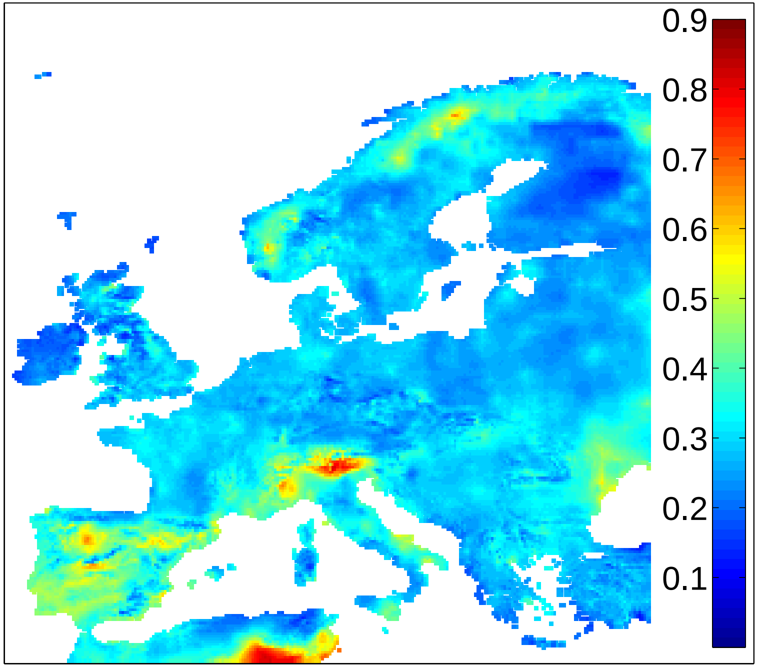

Fig. 2. Coefficient of variation of the winter precipitation (DJFM).

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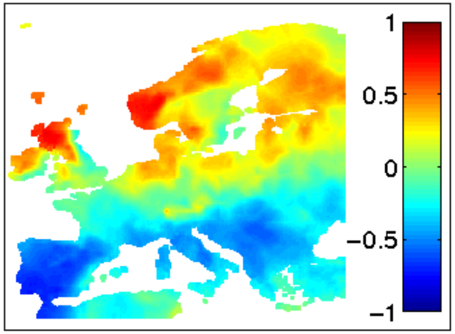

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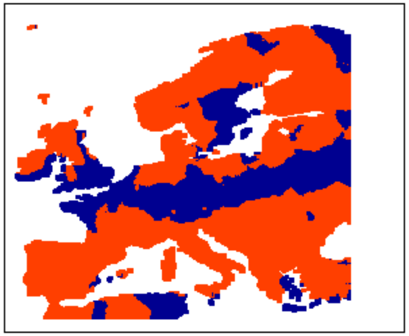

b)

Fig. 3. **(a)** map of Spearman rank correlation coefficients between winter precipitation and North Atlantic Oscillation Index; **(b)** red areas represent regions where the correlation is significant.

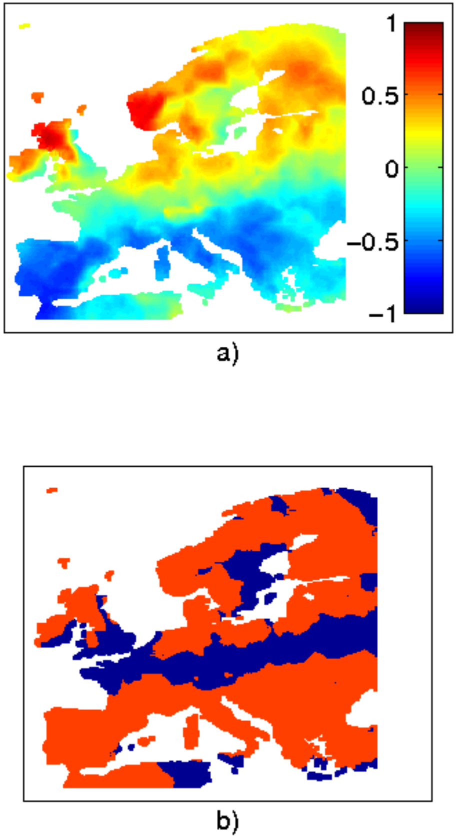

Fig. 4. (a) map of Spearman rank correlation coefficients between winter SPI with modified time scale and North Atlantic Oscillation Index; **(b)** red areas represent regions where the correlation is significant.

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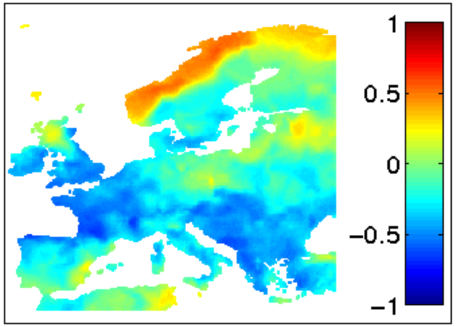

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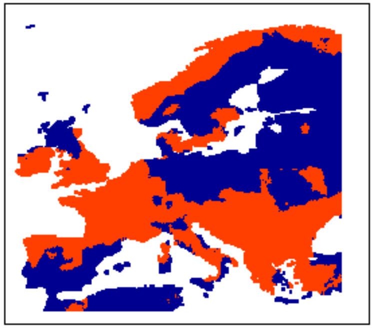

b)

Fig. 5. **(a)** map of map of Spearman rank correlation coefficients between SPI3 of February and East Atlantic West Russia Index; **(b)** red areas represent regions where the correlation is significant.

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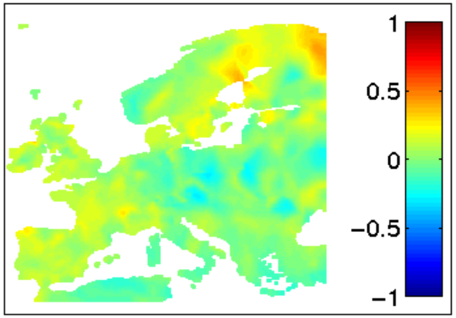

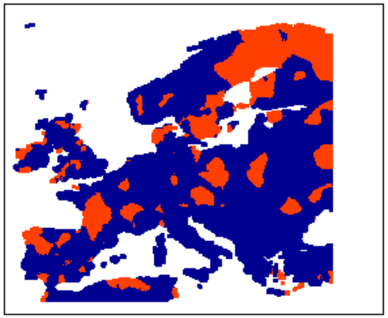

Fig. 6. (a) map of the Mann-Kendall trend coefficients for the winter precipitation; **(b)** red areas represent regions where a significant trend has been detected.

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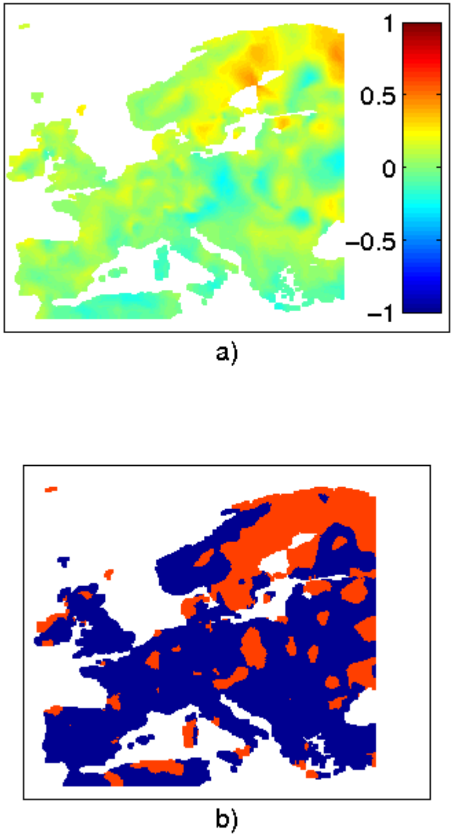

Fig. 7. **(a)** map of the Mann-Kendall trend coefficients for the mean winter SPI3; **(b)** red areas represent regions where a significant trend has been detected.

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