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European precipitation connections with large-scale mean sea-level pressure (MSLP) fields

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Abstract To advance understanding of hydroclimatological processes, this paper links spatiotemporal variability in gridded European precipitation and large-scale mean sea-level pressure (MSLP) time series (1957–2002) using monthly concurrent correlation. Strong negative (positive) correlation near Iceland and (the Azores) is apparent for precipitation in northwest Europe, confirming a positive North Atlantic Oscillation (NAO) association. An opposing pattern is found for southwest Europe, and the Mediterranean in winter. In the lee of mountains, MSLP correlation is lower reflecting reduced influence of westerlies on precipitation generation. Importantly, European precipitation is shown to be controlled by physically interpretable climate patterns that change in extent and position from month to month. In spring, MSLP–precipitation correlation patterns move and shrink, reaching a minimum in summer, before expanding in the autumn, and forming an NAO-like dipole in winter. These space–time shifts in correlation regions explain why fixed-point NAO indices have limited ability to resolve precipitation for some European locations and seasons.

Key words large-scale climate; precipitation; Europe; ERA-40; NAO; hydroclimatological variability

Relations entre les précipitations en Europe et les champs de pression au niveau moyen de la mer à grande échelle

Résumé En vue de faire avancer notre compréhension des processus hydro-climatiques, cet article établit des liens entre la variabilité spatio-temporelle des précipitations européennes (distribuées selon un maillage régulier) et la pression atmosphérique au niveau moyen de la mer PNMM (période 1952–2002) en utilisant les corrélations existant entre elles. Il s'agit pour les précipitations du Nord-Ouest de l'Europe d'importantes corrélations, négatives près de l'Islande et positives près des Açores, ce qui confirme un lien direct avec l'Oscillation Nord Atlantique (ONA). On trouve une structure inverse pour le Sud-Ouest de l'Europe, et la Méditerranée en hiver. Du côté abrité des montagnes, les corrélations avec la PNMM sont plus faibles, reflétant une moindre influence des vents d'Ouest sur la génération des précipitations. Il est important de noter que les précipitations européennes sont contrôlées par des structures climatiques ayant une interprétation physique et qui changent en étendue et position d'un mois à l'autre. Au printemps, les structures de corrélation entre la PNMM et les précipitations se déplacent et s'amenuisent, atteignant un minimum en été, avant de croître en automne, et de former un dipôle de type ONA en hiver. Les déplacements dans l'espace et le temps des régions de corrélation expliquent pourquoi des indices fixes de type ONA ont, en Europe, une aptitude limitée à expliquer les précipitations de certaines régions et de certaines saisons.

Mots clefs climat à grande échelle; précipitations; Europe; ERA-40; ONA; variabilité hydro-climatique

1 INTRODUCTION

European precipitation receipt is dynamic seasonally and spatially (Zveryaev 2004). This variability in precipitation can lead to floods or droughts, which have major socio-economic impacts over Europe (Lorenzo *et al.* 2008, Zveryaev and Allan 2010). Agriculture, water resources management and other sectors are reliant on timely and sufficient precipitation supply, with extreme variability potentially causing water shortages and crop failures, or flood inundation in both urban and rural areas. Identification of hydroclimatological relationships between large-scale climatic circulation and precipitation occurrence helps in understanding the climate drivers of precipitation, which may lead to an improvement in the skill of climate prediction (Zveryaev and Allan 2010). With skilful climate prediction, it would be possible to anticipate precipitation and associated hydrological anomalies, which would help mitigate negative impacts and provide societal benefits.

The average pole-to-equator temperature gradient in the Northern Hemisphere is larger in winter than summer. According to the "thermal wind relationship", a horizontal temperature gradient causes vertical zonal (west–east) wind shear resulting in a region of maximum zonal wind near the tropopause that is called the "jet stream" (Holton 1992). As the largest temperature gradient occurs in winter, jet streams are also strongest in winter. Over the North Atlantic, they occur just east of North America between 30°N and 35°N in winter and between 40°N and 45°N in summer. Extra-tropical cyclones develop in these jet stream regions and travel eastward along storm tracks towards Europe (Holton 1992). In winter, extra-tropical cyclones pick up moisture from the North Atlantic and transport the warm moist air and precipitation over Europe. Recent research has shown that heavy precipitation and flooding are related to "atmospheric rivers", narrow regions of enhanced moisture transport within extra-tropical cyclones (Lavers *et al.* 2011). Regions in northern Europe (e.g. British Isles and Scandinavia) are dominated by westerly winds throughout the year, and are thus most affected by extra-tropical cyclones; the Mediterranean also experiences these storms in the winter, but is influenced by the Subtropical High Pressure Belt in summer. Inland areas (e.g. Central Europe) are less affected by extra-tropical cyclones from the North Atlantic Ocean (Wibig 1999), and so experience a continental climate (Berg *et al.* 2009). In turn, high winter precipitation is found over Western Europe which is intensified by coastal mountains that force the moisture-laden Atlantic air to rise (i.e. orographic enhancement). In contrast, low winter precipitation occurs over Eastern Europe and Russia. In summer (warm-season), smaller-scale processes, such as convection (i.e. thunderstorm activity), are thought to play a key role in precipitation receipt, in part because European precipitation has a significant statistical relationship with European land surface evaporation (Zveryaev and Allan 2010). Convective precipitation events are more prevalent in summer (Berg *et al.* 2009), with precipitation generally being larger over the central continental parts of Europe and lower near the continental extremities (Zveryaev 2004). Interestingly, summer evaporation from the North Atlantic is not found to relate to European precipitation, which is in stark contrast to the winter (Zveryaev and Allan 2010). This suggests that the strong winter jet stream and associated extra-tropical cyclones subside in summer (because of the weaker pole-to-equator temperature gradient), thus limiting this mode of conveyance of North Atlantic moisture and precipitation over Europe.

Indices that describe the state of the large-scale climatic circulation have frequently been used to quantify the connection between the large-scale atmosphere and precipitation (as reviewed below). The North Atlantic Oscillation (NAO), which refers to the redistribution of atmospheric mass between the subtropical Atlantic and the Arctic, has been considered to be the leading mode of atmospheric variability in the North Atlantic basin (Marshall *et al.* 2001), although research suggests that it may not be the most important feature behind precipitation receipt (Qian *et al.* 2000). Changes in the NAO phase, as characterized by the NAO index (NAOI), are associated with variations in the frequency and strength of the surface westerly winds over Europe, influencing the transport and convergence of atmospheric moisture and in turn, regional precipitation patterns (Marshall *et al.* 2001). Winter Scandinavian and Baltic precipitation increases with westerly winds, and hence for a positive NAOI (Hurrell 1995, Uvo 2003, Jaagus *et al.* 2010). Precipitation in northern Britain has also shown a significant positive correlation with the NAOI in winter (Wilby *et al.* 1997, Fowler and Kilsby 2002). Conversely, over the Iberian Peninsula, winter precipitation is negatively correlated with the NAOI (Hurrell 1995, Lorenzo *et al.* 2008). Brandimarte *et al.* (2011) also found a negative correlation between the NAOI and hydroclimatic variables in southern Italy. In the European Alps, precipitation has shown little relation with the NAOI, possibly due to the complex terrain (Bartolini *et al.* 2009). Furthermore, the NAOI has been shown to have a significant influence on extreme winter precipitation (Haylock and Goodess 2004). In general, the NAOI has a stronger link with winter precipitation in coastal European countries, such as Greece, Spain, parts of Scandinavia and the UK (Bouwer *et al.* 2008), and a weaker link with winter precipitation in European regions more remote from the Atlantic Ocean (Wibig 1999). The Scandinavian pattern (Barnston and Livezey 1987) has also been used in previous analyses. A positive phase of the Scandinavian pattern has an anticyclone over Scandinavia and weaker cyclonic regions over Western Europe and eastern Russia, which results in lower precipitation totals over Scandinavia and higher precipitation totals over Central Europe. For further details and maps of Northern Hemisphere teleconnection patterns, the authors refer readers to the National Centers for Environmental Prediction (NCEP) Climate Prediction Center (CPC) website (http://www.cpc.ncep.noaa.gov/data/teledoc/telecontents.shtml).

As the atmosphere is most dynamically active during winter, most previous analyses have focused on the winter season (Folland *et al.* 2009) and the NAOI, with far less attention given to investigating the large-scale climatic circulation and its impacts in summer (Zveryaev 2004, Zveryaev and Allan 2010). In summer, there is a leading pattern of climatic variability known as the summer North Atlantic Oscillation (SNAO) pattern. The SNAO is spatially smaller and located further north than the well-known NAO definition. In a positive SNAO phase, high pressure is present over northwest Europe and low pressure is present over Greenland and the Mediterranean (Zveryaev 2004). Under this circulation pattern, warm and dry conditions occur over northwest Europe (e.g. British Isles), and cool and wet conditions occur over southern Europe and the Mediterranean. As such, precipitation has significant negative correlation with the SNAO in northwest Europe, and significant positive correlation with the SNAO in southern Europe (Folland *et al.* 2009). Historical observations have shown there to be significant statistical relationships between the large-scale circulation and precipitation, but only for some European locations and seasons. Few studies have investigated systematically the important large-scale atmospheric influence on summer precipitation by month at the continental scale.

The availability of gridded observed precipitation and gridded atmospheric re-analysis mean sea-level pressure (MSLP) data has made it possible to undertake a consistent and systematic spatiotemporal analysis of the large-scale climatic control on European precipitation. The aim of this paper is to understand the spatiotemporal variability of European precipitation by quantifying the connections with large-scale MSLP fields throughout the year. This will reveal where and when the hydroclimatological links are strongest, and give insight into the different precipitation-generating atmospheric circulations across Europe. This paper is structured as follows: Section 2 contains the data and methodology; Section 3 discusses the precipitation variability and the field significance and trends in the results; Section 4 presents the relationships between MSLP and precipitation in gridded data sets across Europe and Section 5 examines the connection between the NAOI and precipitation in Europe. Section 6 draws conclusions on the results found.

2 DATA AND METHODOLOGY

Two gridded data sets were used: (a) daily observed gauge-based precipitation (E-OBS version 1.1 data set) produced by the ENSEMBLES project (Hewitt and Griggs 2004, Haylock *et al.* 2008) at a 0.5° × 0.5° resolution across Central and Western Europe (36.25° N–74.25°N; 10.25°W–24.75°E) (North African precipitation time series in the study domain were not used as some series were incomplete); and (b) daily MSLP from the ERA-40 re-analysis data set (Uppala *et al.* 2005) on a 2.5° × 2.5° grid over half of the Northern Hemisphere (0°N–90°N; 90°W–90°E). The MSLP is used as an explanatory variable of European precipitation, as MSLP is co-located with vertical velocity in the mid-troposphere and, therefore, indicative of cyclonic development and precipitation. Time series of monthly precipitation and MSLP were derived for the common data period of September 1957–August 2002. Monthly data were used in our analysis because daily data have been found to be too noisy to detect the fundamental climatic controls on precipitation (Lavers *et al.* 2010). Two versions of the NAOI were used to act as a benchmark against which to compare the links with gridded MSLP: (1) a fixed station-based NAOI from the University of East Anglia (UEA) Climatic Research Unit (CRU) (http://www.cru.uea.ac.uk/cru/data/nao/nao.dat), and (2) an Empirical Orthogonal Function (EOF) based NAOI from the Climate Analysis Section at the National Center for Atmospheric Research (NCAR) (http://www.cgd.ucar.edu/cas/jhurrell/indices.data.html).

The univariate normality of the MSLP and precipitation time series was tested using the Shapiro-Wilk test (significance level $\alpha = 0.05$) as it is one of the most powerful tests for detecting non-normality (Helsel and Hirsch 1992). Results suggested that some time series were not normally distributed, which means that it is not appropriate to use a linear Pearson correlation coefficient as a measure of association. Therefore, the non-parametric Spearman Rank correlation method (ρ) was used herein (significance level $\alpha = 0.05$) (Spearman 1904). Correlation analysis was carried out between MSLP at each grid in the atmospheric domain and each grid of observed precipitation by month ($n = 45$) to assess the detailed spatial variation of MSLP control on European monthly precipitation. In sections 3 and 4 the results are presented for the four seasons (winter, spring, summer and autumn), with January, April, July and October generally being used to represent the seasons respectively.

Monthly field significance of the observed MSLP–precipitation correlation fields (at each precipitation grid) was examined using Monte Carlo simulations to determine whether the observed significant MSLP correlation areas were greater than those expected by chance alone (Livezey and Chen 1983, Phillips and McGregor 2002). Livezey and Chen (1983) suggest using 200 simulations to estimate accurately the probability density function of the number of significant MSLP–precipitation correlations observed in each simulation. For each simulation, a series of 45 values (i.e. same as the length of the monthly time series over the period 1957–2002) was generated randomly from the empirical distribution of the precipitation time series, and then correlated with the MSLP time series at each of the 2701 grid points in the MSLP field. The number of MSLP grid points with significant ranked correlation at the 0.05 level was recorded. An observed correlation pattern is considered field significant at the 0.05 level, if the area of observed significant correlation is larger than that expected by chance, as given by the 95% percentile of the empirical probability distribution constructed from the 200 Monte Carlo simulations. Correlation patterns that are field significant are to be considered as possible centres of climatic circulation related to precipitation.

The Mann-Kendall trend test (significance level $\alpha = 0.05$) was used to determine if trends were present in the monthly MSLP and precipitation time series (e.g. Helsel and Hirsch 1992) and to assess if spurious MSLP–precipitation correlations could result as a result of any trends. For each month, precipitation time series that had increasing or decreasing linear trends (at $\alpha = 0.05$) underwent a bootstrap procedure to determine if the presence of the trend affected the significance of the correlations with MSLP (following Efron and Tibshirani (1998)). Note that the bootstrapping process destroys any temporal trends in both the MSLP and precipitation time series. The following bootstrap process was repeated $B = 1000$ times. For each bootstrap sample B, precipitation (at a particular grid) and MSLP (at all 2701 grid points) time series of the same size as the observed series ($n = 45$) were re-sampled with replacement keeping the concurrent pairs. At each MSLP grid cell, Spearman's correlation ρ between re-sampled precipitation (at a particular grid) and MSLP was computed, providing an empirical bootstrap distribution of 1000 ρ values for each MSLP grid point. If 95% of the constructed empirical bootstrap distribution has a correlation $\rho > 0$ or $\rho < 0$, the trend has no impact on the significance of the correlation at the 0.05 significance level. This significance threshold was used for all MSLP grid points for the relevant precipitation grids. Using this approach, it was possible to assess if the correlations obtained for the observed time series were significant despite a possible presence of trends in these time series.

A spatially nested research design is adopted to facilitate clear presentation of results for different spatial scales and locations. Six precipitation grid locations were chosen *a priori* (to give a large spread across Europe) to test the hypothesis that precipitation dynamics are different across Europe. These locations are: (a) western Scotland (57.25°N, 5.25°W), (b) Norway (65.25°N, 13.75°E), (c) southern Spain (37.75°N, 3.75°W), (d) central France (47.25°N, 3.25°E), (e) Czech Republic (49.75°N, 15.25°E) and (f) the Balkans (42.75°N, 20.25°E). Scaling-up from these individual grids, results are presented for all precipitation grids across the British Isles and, at a larger scale still, for all precipitation grids across Europe. The latter provides a novel wider perspective on connections between MSLP and precipitation in Europe.

3 PRECIPITATION VARIABILITY, FIELD SIGNIFICANCE TESTING AND TREND ANALYSIS

3.1 European precipitation variability and the CRU NAOI

Figure 1 shows the monthly standardized time series of precipitation anomalies for the six locations across Europe (referring now to the given six

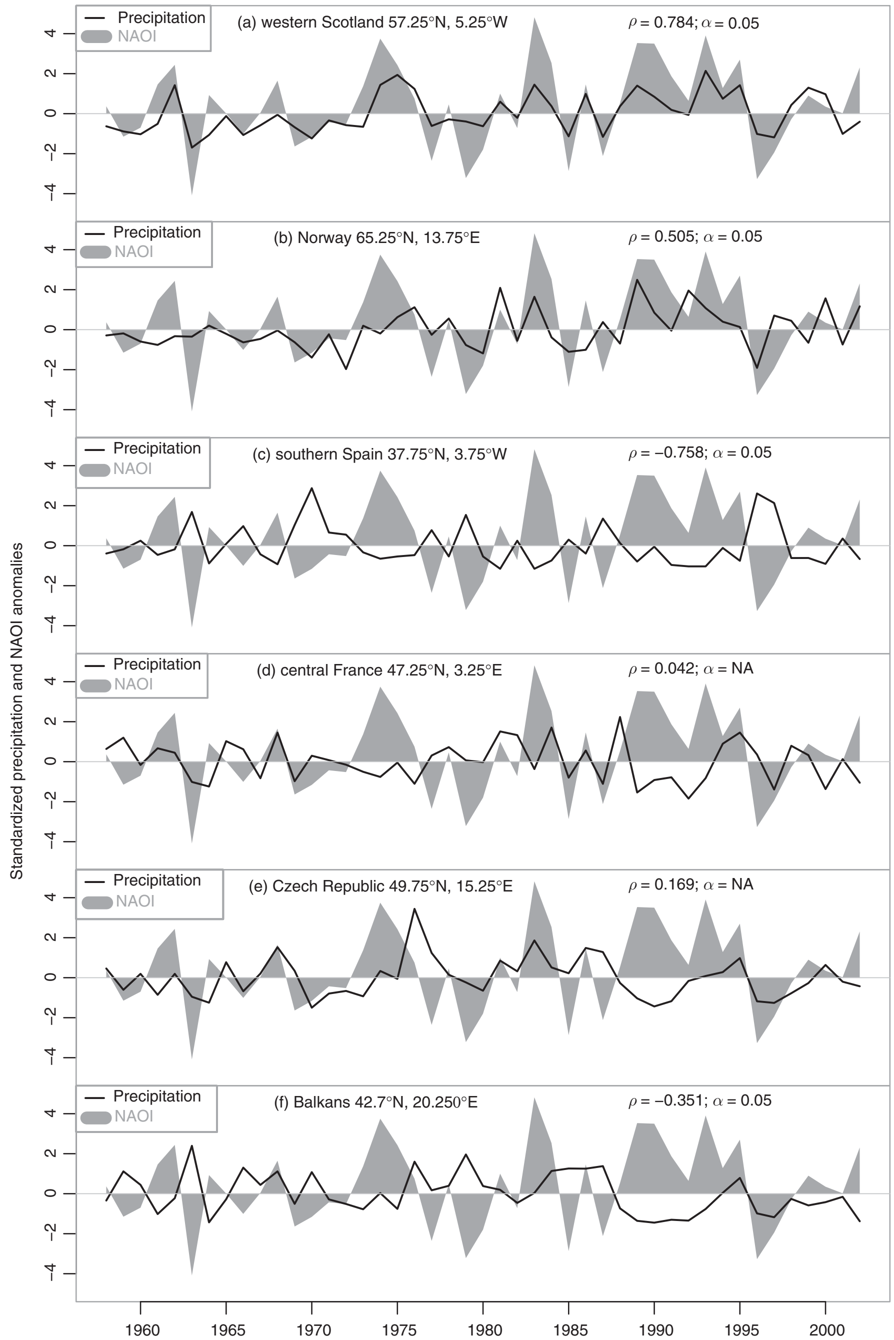

Fig. 1 Monthly standardized time series of January precipitation anomalies in six grid point locations across Europe and the station-based CRU NAOI (1958–2002). The areas shown are: (a) western Scotland (57.25°N, 5.25°W), (b) Norway (65.25°N, 13.75°E), (c) southern Spain (37.75°N, 3.75°W), (d) central France (47.25°N, 3.25°E), (e) Czech Republic (49.75°N, 15.25°E), and (f) Balkans (42.75°N, 20.25°E). The CRU NAOI–precipitation Spearman correlations are given for each of the six precipitation grids with the significance level α (NA implies the correlation is not significant at the 0.05 level). The precipitation time series are solid black lines and the CRU NAOI time series are shaded solid grey.

locations) along with the CRU NAOI series (NAOI–precipitation correlation analyses are the focus of Section 5). In general, precipitation anomalies are in phase with the NAOI in northern Europe (western Scotland and Norway; Fig. 1(a) and (b), respectively) and out of phase with the NAOI in southern Europe (southern Spain and the Balkans; Fig. 1(c) and (f), respectively), which is consistent with the regional precipitation linkages with the large-scale climatic circulation (Hurrell 1995). There is less inter-regional agreement between the precipitation time series in July (Fig. 2), possibly indicating more local-scale

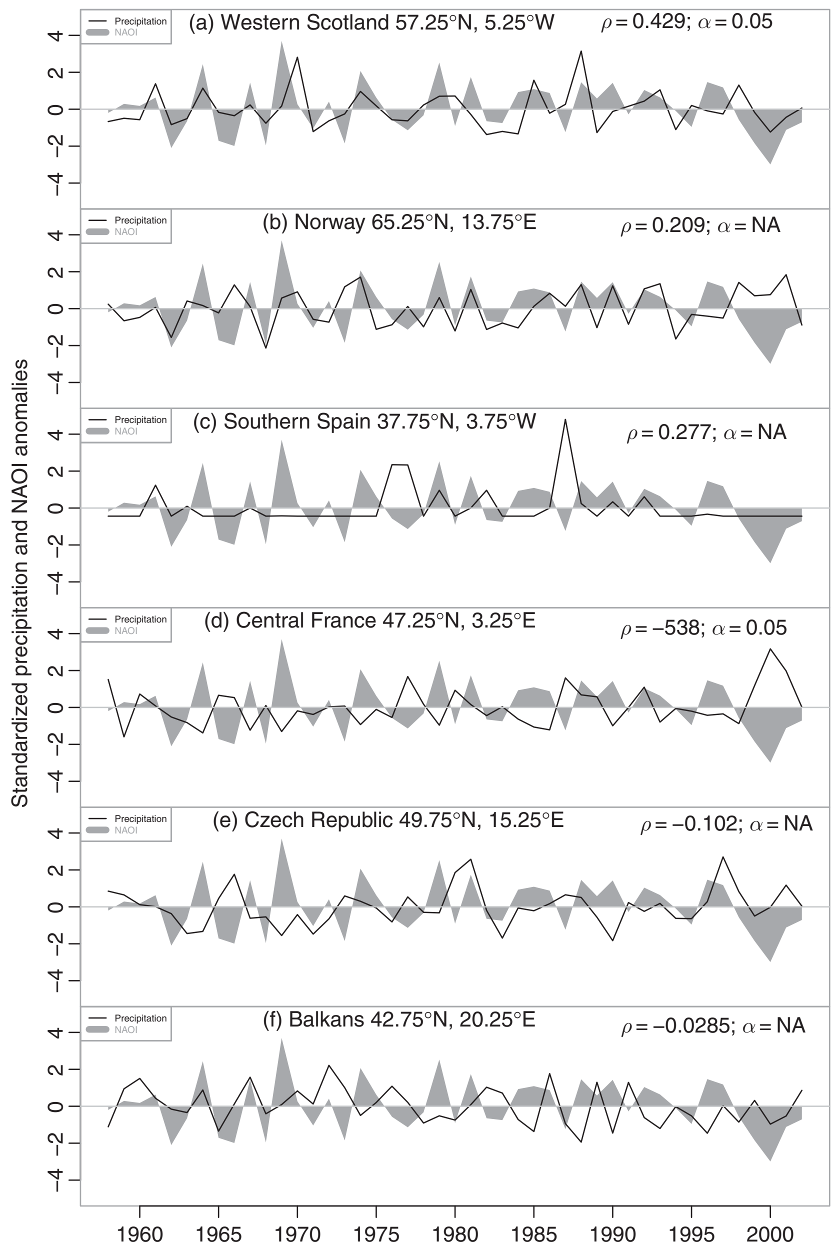

Fig. 2 Monthly standardized time series of July precipitation anomalies (1958–2002) in six grid point locations across Europe and the station-based CRU NAOI (1958–2002) [Key as Fig. 1].

convective precipitation generating processes in summer (Berg *et al.* 2009); an exception is some noticeable co-variability between precipitation time series in western Scotland (Fig. 2(a)) and Norway (Fig. 2(b)). The lack of precipitation variability in southern Spain (Fig. 2(c)) reflects the high number of years with no July rainfall, a climatic feature shared amongst a large area surrounding the Mediterranean.

3.2 Field significance

Field significance was assessed to determine if the area (number of grid cells) with observed significant monthly correlation between MSLP and precipitation was larger than what would be expected by chance alone. If this is the case, the month was described as "field significant". The number of months in each season with field-significant correlations between precipitation and MSLP is shown in Fig. 3. In winter (December, January and February; DJF) in Fig. 3(a), coastal European regions (except France) have the largest number of field-significant correlation patterns (e.g. Balkans, Iberian Peninsula and western Scandinavia) suggesting that the large-scale climatic circulation has a significant influence on coastal European precipitation in winter; this corroborates

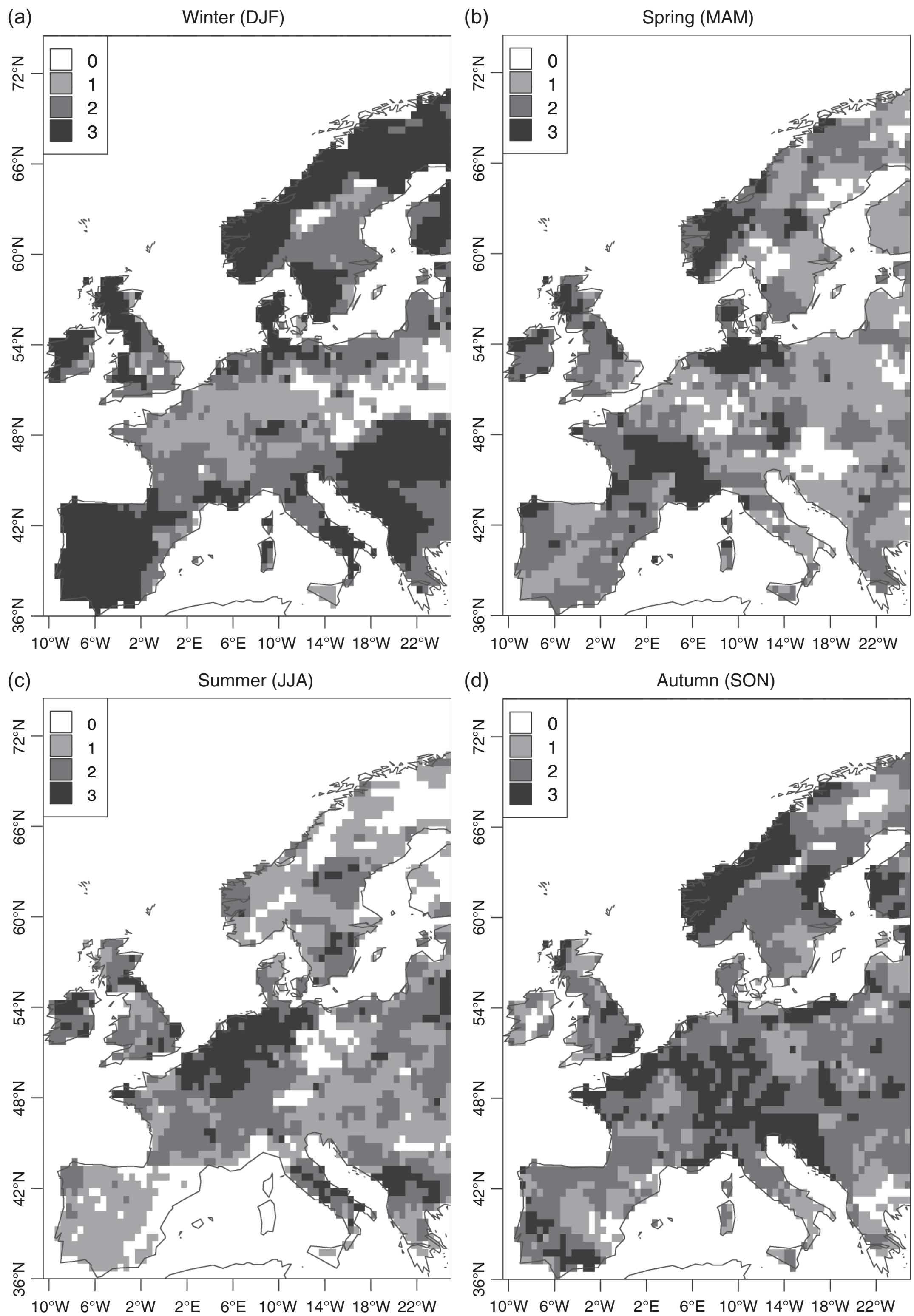

Fig. 3 Number of field significant months for MSLP for: (a) winter (DJF), (b) spring (MAM), (c) summer (JJA) and (d) autumn (SON) (period of study is September 1957 to August 2002).

the findings of Wibig (1999). With increasing distance from the Atlantic Ocean and in the lee of mountain ranges (central and east Sweden, central Britain and in the lee of the European Alps), the number of field significant months decreases because of weaker connection between precipitation and the large-scale atmosphere (Fig. 3(a)). In spring (March, April and May; MAM) and autumn (September, October and November; SON) in northwest Britain and western Scandinavia, the many months with field significant precipitation grids may indicate that the winter climatic circulation exists for a longer duration (Fig. 3(b) and (d), respectively). Fewer field significant precipitation grids were found in summer (Fig. 3(c)) suggesting that more local-scale weather systems produce the precipitation in summer, because the observed significant MSLP–precipitation correlation areas exist across smaller geographical regions. The field significant months over the Low Countries (defined as Belgium, The Netherlands, Luxembourg and parts of northern France and western Germany) and Italy/Balkans in July (Fig. 3(c)) indicates a relationship with the SNAO pattern.

3.3 Influence of trends on the correlation analyses

The precipitation and MSLP time series were tested for trends that may affect the significance of the correlations. Results of the Mann-Kendall test suggest that trends in precipitation time series occur predominantly in winter (January–March). In Scandinavia (Spain), precipitation has an increasing (a decreasing) trend over 1958–2002. Around 45°N (i.e. the Alps) is the transition zone between increasing and decreasing trends. During winter, significant increasing (decreasing) MSLP trends are found near the Azores (Iceland), indicating a tightening of the pressure gradient over the North Atlantic and a stronger zonal (westerly) flow, coincident with a stronger positive phase of the NAO. This is consistent with Hurrell and Van Loon (1997), who found that the NAO was often in a positive phase from 1980 until the late 1990s. An increasingly stronger positive NAO could be the cause of the significant increasing (decreasing) precipitation trend observed over Scandinavia (the Iberian Peninsula). Furthermore, the trend towards a strengthening of the NAO influence on European climate during the last part of the 20th century is thought to be due to an eastward movement (i.e. towards Europe) of MSLP anomalies associated with the NAO (Vicente-Serrano and Lopez-Moreno 2008).

The influence of trends (in both the precipitation and MSLP time series) on the significance of the correlations was assessed using a bootstrap procedure. Figure 4 shows the empirical bootstrap distribution of the 1000 correlations produced between the

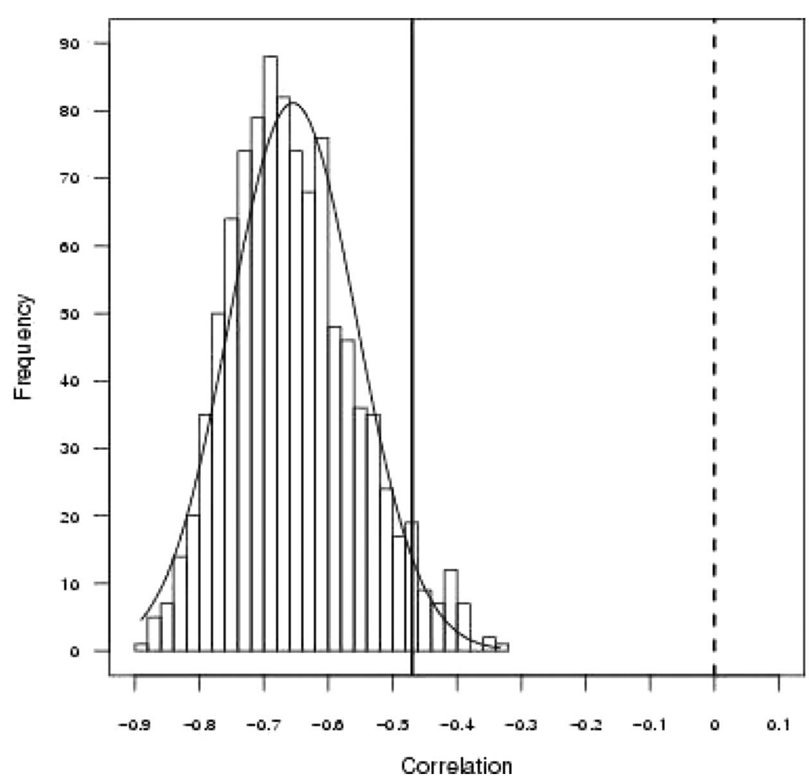

Fig. 4 The empirical bootstrap distribution of the correlation from the 1000 realizations between the February precipitation time series with the strongest increasing trend (62.25°N, 9.75°E; Norway) and the MSLP time series (70°N, 7.5°W) with the highest observed correlation with precipitation. Solid vertical line is the 95% percentile; dashed line is correlation $\rho = 0$.

grid point with precipitation time series with the strongest increasing trend (February; 62.25°N, 9.75°E; Norway) and the MSLP grid point with the strongest observed correlation in February ($\rho =$ –0.667; 70°N, 7.5°W). The 95% percentile of the empirical bootstrap correlation distribution (black vertical line in Fig. 4) is less than zero; therefore, the trend does not affect the significance of the observed correlation at this grid point at the 0.05 significance level. Hence, the correlation is significant even without a trend present. Results of the empirical bootstrap distributions at all MSLP grid points for the precipitation grids with significant trends reveal that the significance levels (i.e. 0.05) were almost identical to those obtained through the observed correlation analysis, demonstrating more widely that the presence of trends does not significantly affect the results presented herein.

4 CORRELATION BETWEEN MSLP AND PRECIPITATION

This section presents the correlation analyses of large-scale climatic circulation (MSLP) with European precipitation. A precipitation grid point in western Scotland (57.25°N 5.25°W) is used to illustrate how the correlation results are presented because of this location's closeness to the westerly flow that results in strong MSLP–precipitation relationships. Figure 5 shows the map of correlations in January between the precipitation time series in western Scotland and the gridded MSLP field across half of the Northern Hemisphere. This map shows a correlation dipole with significant negative correlation (blue colour) centred over the Norwegian Sea and significant positive correlation (red colour) centred near the Azores. This dipole implies that as MSLP falls to the north and rises to the southwest of the British Isles, precipitation increases in western Scotland, which relates to the Icelandic Low and Azores High pressure systems respectively. Note that the continuous blue colour at high latitudes (at 90°N) occurs because there is only one MSLP value (i.e. an artefact of the cartographic projection chosen). Following the nested research design adopted, results (with correlation maps similar to Fig. 5) are presented first for precipitation grid points in the British Isles (Fig. 6) and secondly for Europe (Figs 7–10) for selected months. The results presented in Figs 6–10 cannot be reproduced in a

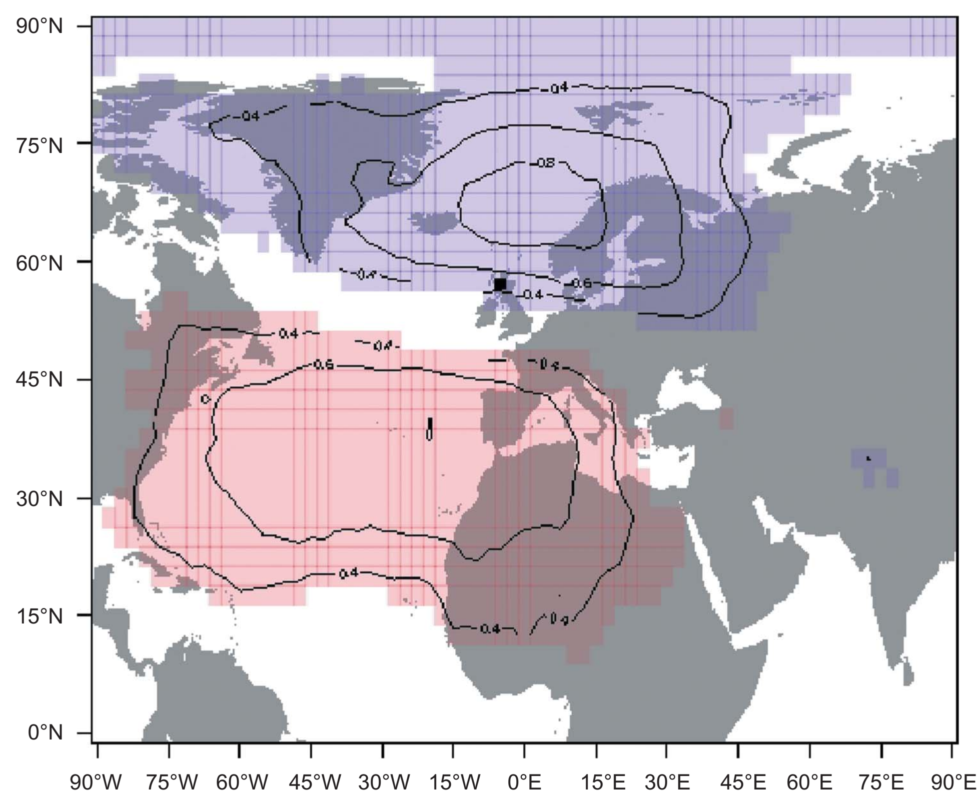

Fig. 5 Correlation analysis of precipitation in a single grid in western Scotland (57.25°N, 5.25°W; location given by black box) with MSLP in January (1958–2002). Shading signifies significant Spearman rank correlation (significance level $\alpha = 0.05$) (cf. Fig 6).

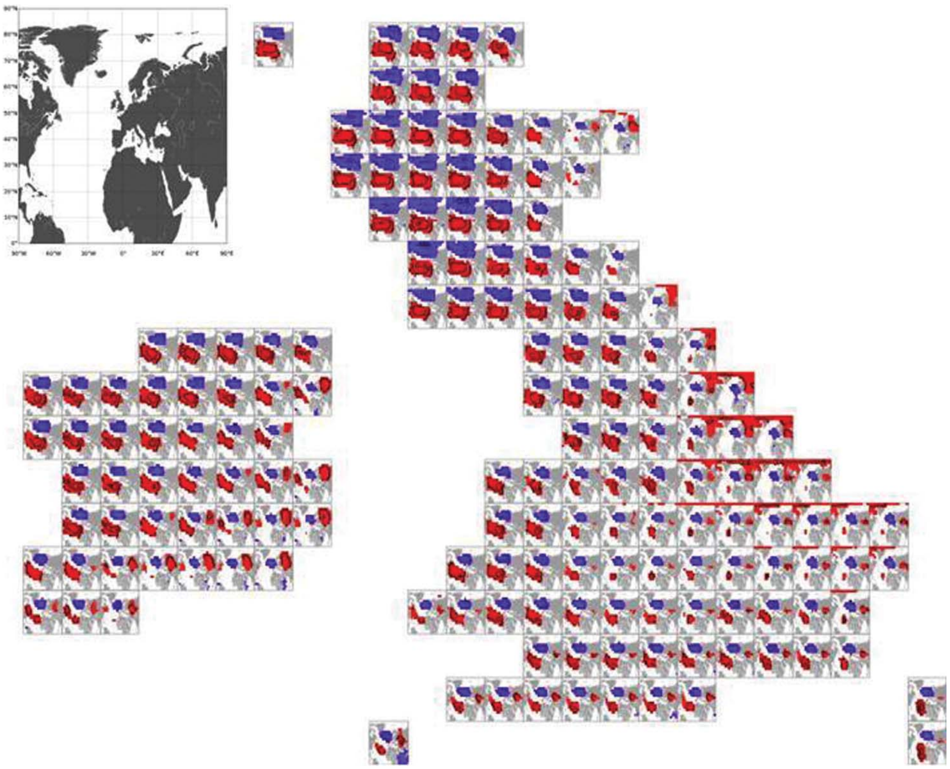

Fig. 6 Correlation analysis of gridded precipitation over the British Isles with MSLP for January 1958–2002. The grey background in each grid box represents the land masses in the study domain and the geographical map shown in the top-left corner represents the spatial domain used. Red (blue) colour signifies significant positive (negative) Spearman rank correlation (significance level a = 0.05).

larger format given the journal page layout. In these diagrams, the main features of interest to readers are the changes in size and location of the colours (i.e. significant correlation patterns). Each grid point covers the region displayed as a map inset (with the same aspect ratio) in Figs 6–10.

4.1 Correlation analysis over the British Isles in January

In January, across the British Isles the location of areas with significant correlation between MSLP and precipitation vary both in size and location (coloured areas in Fig. 6). In western Scotland, the correlation dipole has significant negative correlation to the north and significant positive correlation to the southwest of the British Isles (as in Fig. 5); this relates to the Icelandic Low and Azores High pressure centres, respectively. The large areas of significant correlation between MSLP and precipitation in western British districts (including the western Scottish Highlands, the Pennines and Welsh mountains) are due partly to the orographic enhancement of rainfall (Roy 1997, Sumner 1997, Tufnell 1997). Notably, as distance increases from the Atlantic Ocean, the correlation dipole shrinks and areas with non-significant correlation become more prominent, probably due to the shelter from the westerly air flow by western mountain chains. Precipitation in eastern Britain has no significant positive correlation with MSLP to the southwest of the British Isles. Instead, positive correlation is seen with MSLP over central Russia, which could relate to the Siberian High pressure system (Fig. 6). This suggests that precipitation is related to an easterly air flow on the southern edge of the Siberian High. En route to Britain, the easterly flow would be modified becoming moist due to evaporation over the North Sea, which in turn could produce precipitation in northeast England.

4.2 Correlation analysis over Europe in winter

The MSLP–precipitation correlation dipole seen in January across the western British Isles was also found across northern Norway and Sweden, as shown in Fig. 7 (i.e. across the northwest European boundary with the North Atlantic Ocean). This statistically significant MSLP correlation dipole (ρ >

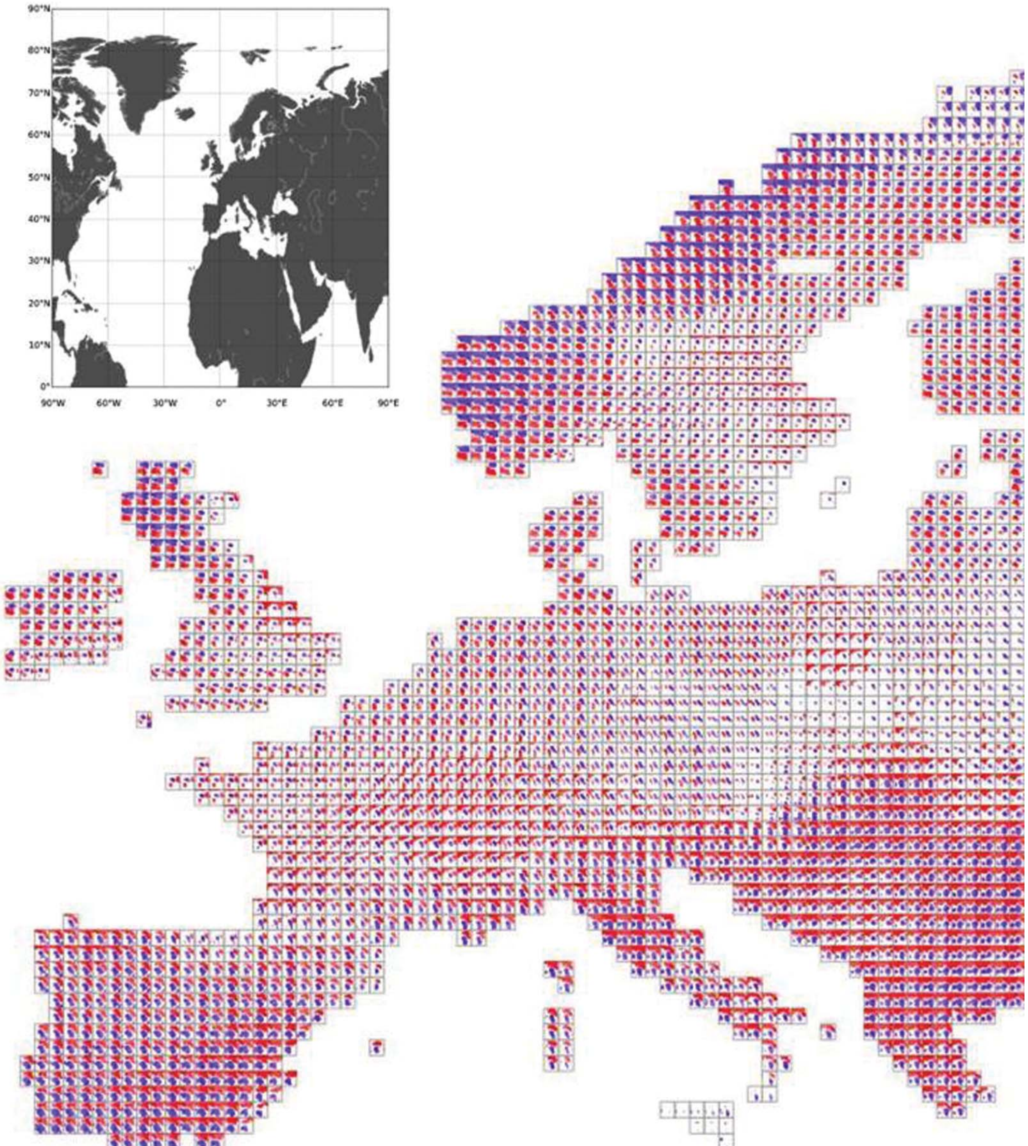

Fig. 7 Correlation analysis of gridded European precipitation with MSLP for January 1958–2002 [Key as Fig. 6].

|0.6|) is strongest in December, January (Fig. 7), and February; and it implies that, from the western British Isles through to Scandinavia, precipitation occurs when MSLP is low near Iceland and high near the Azores. This same dipole structure, although smaller in extent (especially for the positive correlation area), is also found across northern continental Europe from northern France to Finland, indicating that precipitation has a positive relationship with the NAO pattern. These results show that a similar large-scale climatic circulation generates winter precipitation across northern and western Europe.

From western to eastern Scandinavia, the size of significant MSLP–precipitation correlation patterns show a distinct west–east gradient, with stronger and larger MSLP influence on precipitation in the west than in the east. The effect of MSLP on precipitation, as characterized by the correlation patterns, is less in central and east Sweden than in western Norway which is likely to be due to the Scandes Mountains (between Norway and Sweden) reducing the influence of moist westerly winds from the Atlantic on central Sweden (Uvo 2003, Kingston *et al.* 2009). In southern Finland, where there is a lesser influence of mountains on the atmospheric flow, westerly winds can penetrate further into the European continent, thus bringing precipitation (Uvo 2003).

From central France southwards, the winter correlation dipole has a reversed pattern compared to northern Europe (as also identified by Bartolini *et al.*

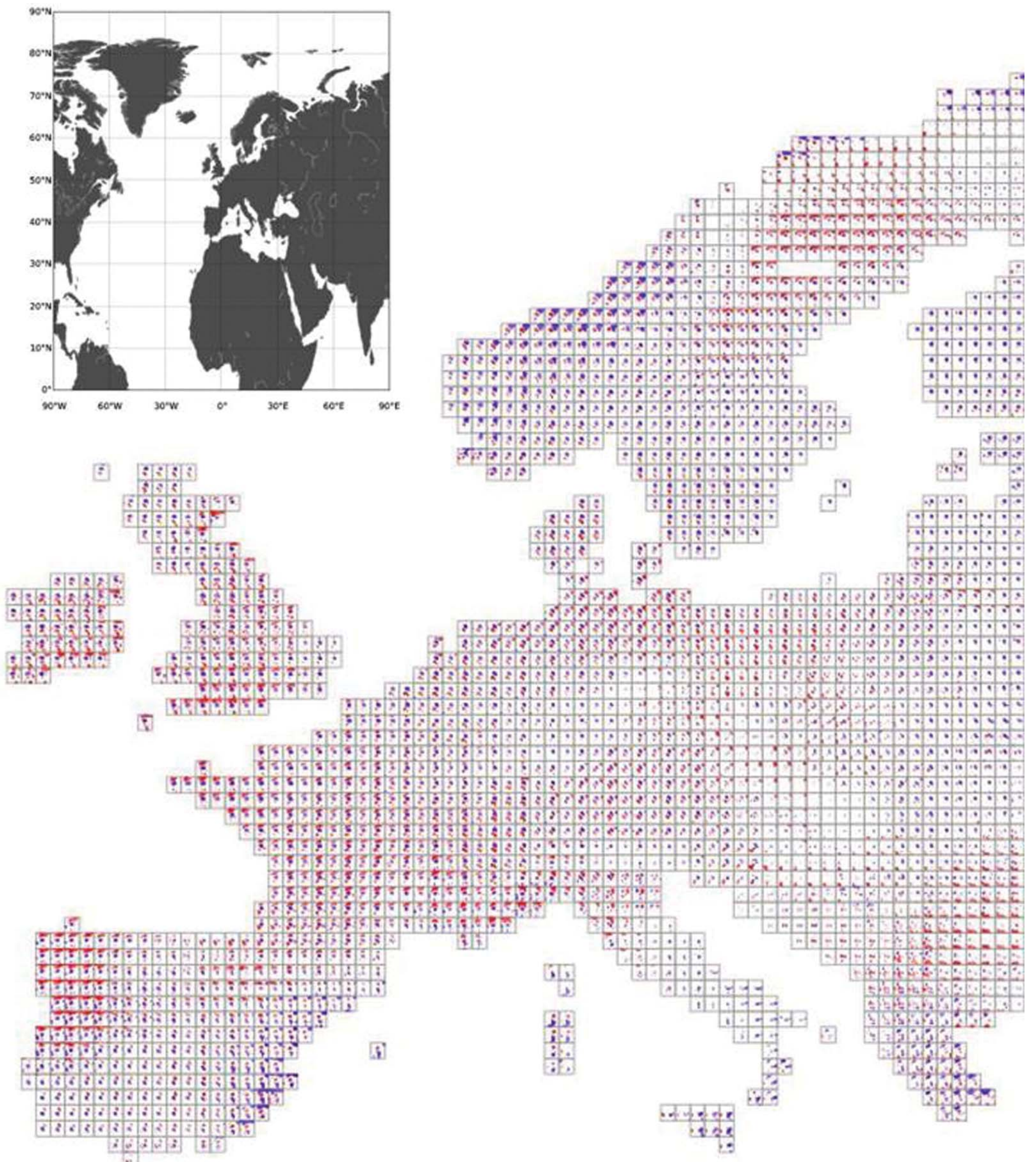

Fig. 8 Correlation analysis of gridded European precipitation with MSLP for April 1958–2002 [Key as Fig. 6].

2009). From the Iberian Peninsula to the Balkans, large regions of negative and positive correlation are centred over the Azores and northern Europe, respectively (Fig. 7). This means that as MSLP falls near the Azores and rises near Iceland, a cyclonic circulation affects southern Europe, thus increasing precipitation. This climatic circulation is associated with storm tracks that are located further south than normal that steer rain-bearing depressions into southern Europe (Marshall *et al.* 2001). In central and eastern Europe downwind of the European Alps, smaller MSLP–precipitation correlation patterns occur indicating that the Alps reduce the influence of westerly winds on precipitation in this region.

4.3 Correlation analysis over Europe in spring

With the onset of spring, the pole-to-equator temperature gradient (hence, the westerly air flow) across the North Atlantic weakens. The MSLP–precipitation correlation patterns for April (Fig. 8) show that smaller-scale climatic circulation patterns (circulation more local to the location where the precipitation is received) are linked with precipitation and the large-scale winter atmospheric patterns have broken down. In southern Britain the correlation dipole has reversed from winter to spring and note how the orientation of the correlation patterns has shifted with the changing season. However, note that in northern Scandinavia

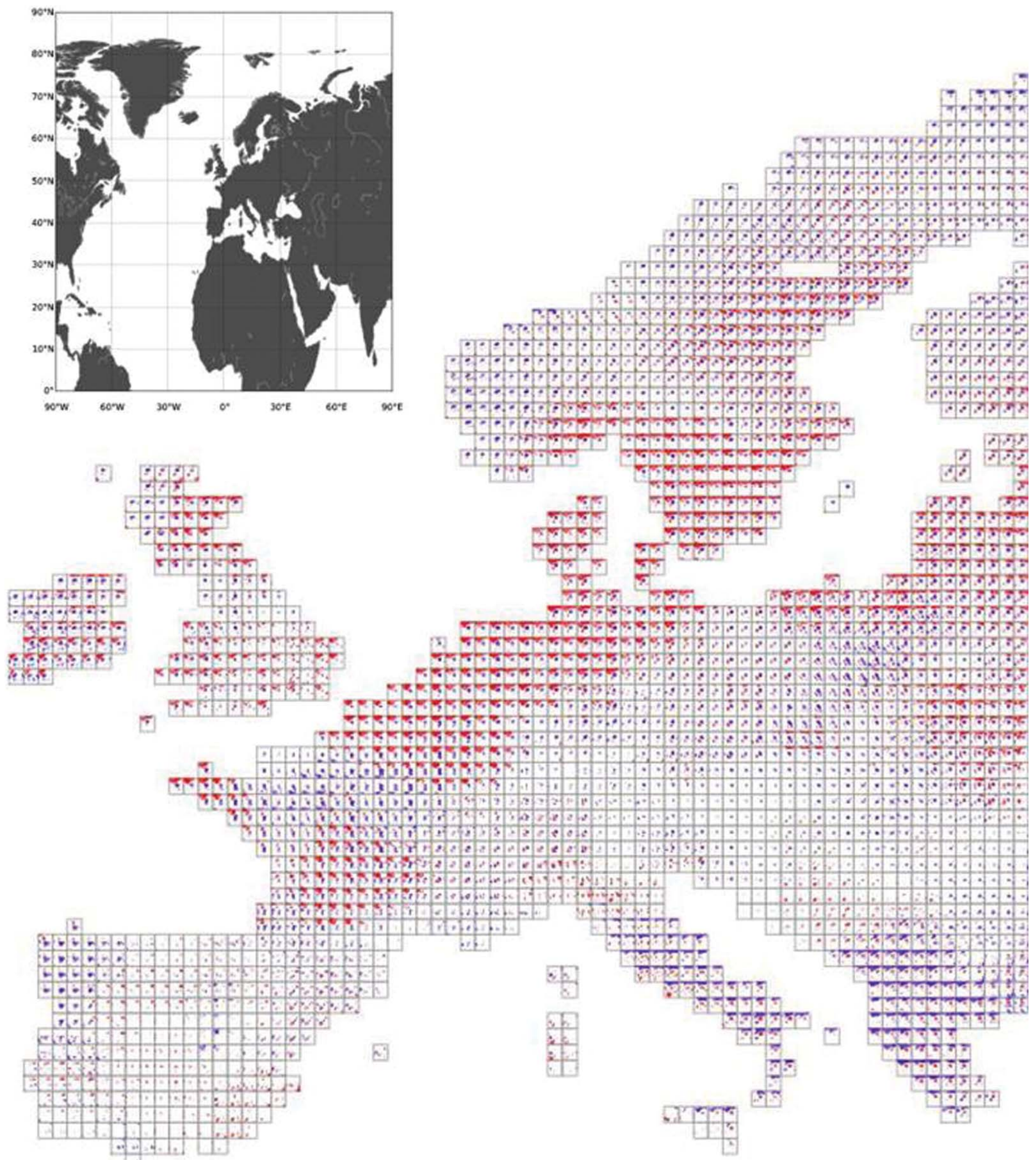

Fig. 9 Correlation analysis of gridded European precipitation with MSLP for July 1958–2002 [Key as Fig. 6].

and in the far northwest of Scotland, a North Atlantic correlation dipole still exists in April suggesting that the winter circulation patterns have a longer duration at higher latitudes.

4.4 Correlation analysis over Europe in summer

As summer approaches, the North Atlantic correlation dipole continues to weaken. During summer (June, July, August; JJA), northern European precipitation has no significant positive correlation over the Azores region. In general, precipitation has fewer significant correlations with the MSLP field (larger white areas in Fig. 9 in July). During June and July, significant positive MSLP correlation over Greenland is found with central-northern European precipitation (e.g. the Netherlands and northern Germany), and significant negative MSLP correlation over Greenland is found with southern to southeast European precipitation (e.g. Greece). This means that as pressure falls over Greenland, pressure rises over northwest Europe resulting in precipitation decrease over northwest Europe; this is shown by the positive (red) correlation in the top-left (northwest) of many precipitation grids in central-northern Europe (Fig. 9). For southern Europe the negative (blue) correlation areas over Greenland (northwest of each precipitation grid) highlight that

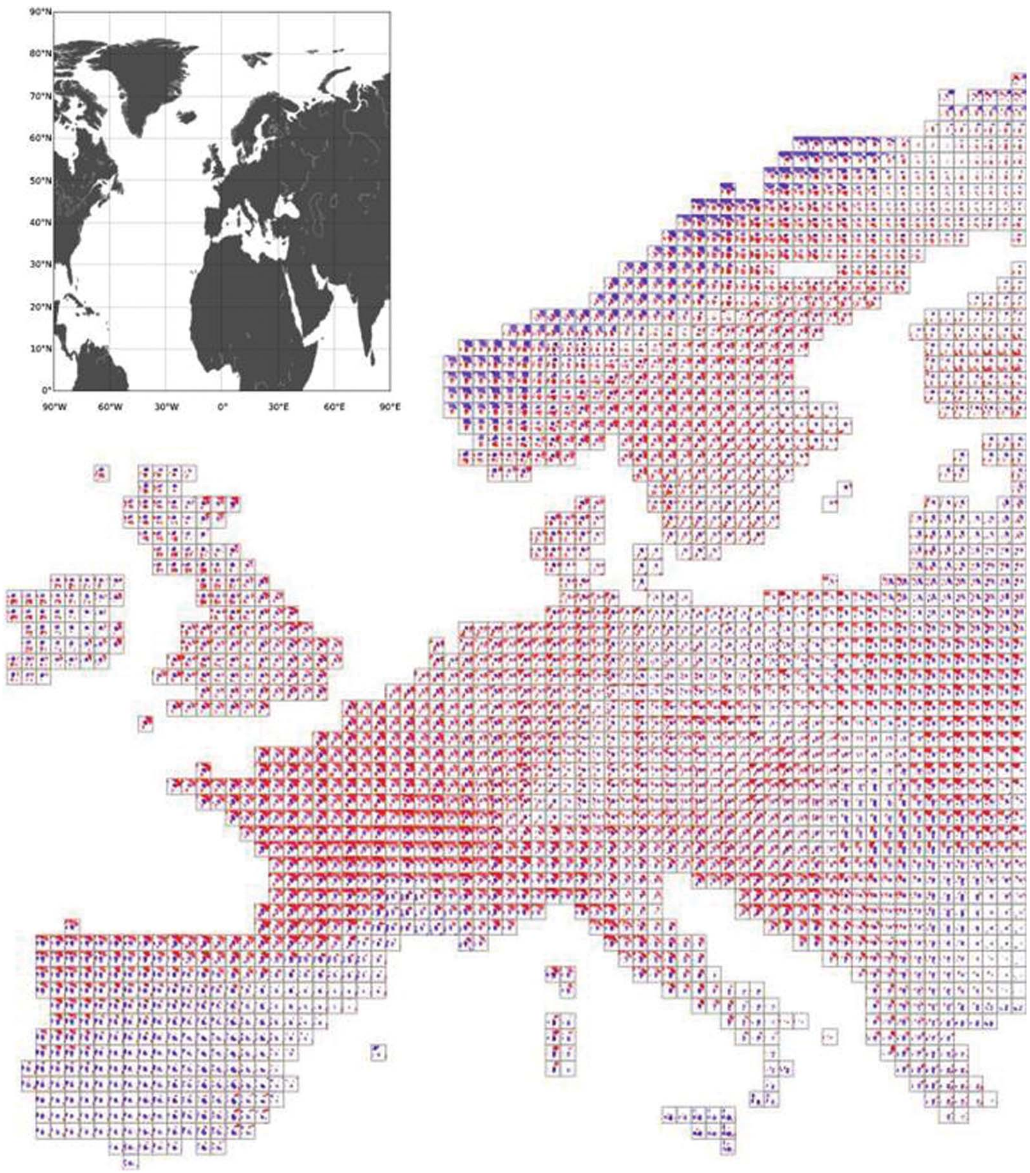

Fig. 10 Correlation analysis of gridded European precipitation with MSLP for October 1957–2001 [Key as Fig. 6].

as pressure falls over Greenland, pressure falls over the Mediterranean, and precipitation increases over southeast Europe. Zveryaev (2004) identified summer relationships between precipitation and patterns of high pressure over northwest Europe and low pressure over Greenland and the Mediterranean; this was described as the SNAO, as discussed in the Introduction. The centres of correlation of MSLP and precipitation shown herein coincide with those of Zveryaev (2004).

The smaller extent of the significant MSLP–precipitation correlation patterns in summer compared to winter suggests that the large-scale climatic circulation has a lesser influence on summer precipitation. This may be related to the prevalence of convective precipitation events in summer (Berg *et al.* 2009) that correspond to smaller-scale synoptic weather systems. Zveryaev and Allan (2010) suggest local processes (i.e. the circulation that causes the precipitation occurs on a smaller scale) play a dominant role in summer precipitation occurrence. In our analysis herein, although the temporal (monthly) and spatial (2.5° × 2.5°) resolution of the MSLP fields are too coarse to resolve convective activity, the small size of correlation patterns identified with European precipitation in summer is a sign of smaller-scale atmospheric dynamics in operation.

4.5 Correlation analysis over Europe in autumn

During autumn, the North Atlantic MSLP–precipitation correlation dipole pattern begins to reappear in September in northern Europe; this is visible in October (Fig. 10) across western British Isles and Scandinavia. Note that over Scandinavia, the MSLP–precipitation correlation patterns have a more easterly location compared to the western British Isles, which suggests a linkage with the Scandinavian climate pattern (Barnston and Livezey 1987); the Scandinavian pattern is discussed further in Section 5. The re-formation of the large-scale North Atlantic climatic circulation is not found until December in the southern British Isles and northern France (not shown) suggesting that the winter climatic circulation pattern is in place earlier further north where the winter season is longer.

5 CORRELATION BETWEEN NAOI AND PRECIPITATION

The maximum and minimum gridded MSLP-precipitation correlations were compared to the precipitation correlations with the CRU NAOI and EOF NAOI for the six selected precipitation time series (as in Figs 1 and 2). Since the NAOI has been used widely as a measure of the strength of the North Atlantic influence on European climate (as reviewed in Section 1), the comparative analysis herein serves to benchmark the gridded MSLP–precipitation correlations. For all six locations, precipitation has stronger correlations with the MSLP than with the NAOI, regardless of season (Table 1 for January; Table 2 for July). This suggests that the NAOI is unable to explain precipitation occurrence as well as MSLP because the centres of strong MSLP correlation do not always coincide with the fixed Azores-Iceland locations defining the NAOI. Note that the CRU NAOI and EOF NAOI provide similar results (Tables 1 and 2). In January (Table 1) northern European precipitation has a positive relationship with the NAOI and southern European precipitation has a negative relationship with the NAOI, as shown by the time series plots in Fig. 1 and as also found by Hurrell (1995). Further east (central France and Czech Republic) the NAOI has lower correlation with precipitation reflecting: (a) a reduced oceanic influence on precipitation, and (b) the areas of climatic control on precipitation are not co-located with the NAOI definition. In July, only western Scotland and central France have significant NAOI–precipitation correlations (Table 2), possibly suggesting that western Scottish precipitation is still related to extra-tropical cyclones that cross the Atlantic Ocean, and that French precipitation is being influenced by the SNAO pattern (Fig. 9).

Other climatic patterns that are characterized by simple indices may also be related to European precipitation. In particular, evidence of the Scandinavian pattern (Barnston and Livezey 1987) can be found in

Table 1 Spearman Rank correlation for the January time series (1958–2002; $n = 45$) of the six precipitation grid points of Fig. 1. Maximum and minimum correlation with the MSLP field and correlation with the station-based CRU NAOI and EOF-based NAOI are shown. Correlation at the significance level $\alpha = 0.05$ is in bold font.

Region	Grid	NAOI (CRU)	NAOI (EOF)	Max. MSLP	Min. MSLP
Western Scotland	57.25°N, 5.25°W	**0.784**	**0.787**	**0.801**	**−0.883**
Norway	65.25°N, 13.75°E	**0.505**	**0.561**	**0.806**	**−0.794**
Southern Spain	37.75°N, 3.75°W	**−0.758**	**−0.703**	**0.740**	**−0.902**
Central France	47.25°N, 3.25°E	0.042	−0.110	**0.566**	**−0.708**
Czech Republic	49.75°N, 15.25°E	0.169	0.086	**0.538**	**−0.630**
Balkans	42.75°N, 20.25°E	**−0.351**	**−0.466**	**0.627**	**−0.842**

Table 2 Spearman Rank correlation for the July time series (1958–2002; $n = 45$) of the six precipitation grid points of Fig. 2 [Key as Table 1].

Region	Grid	NAOI (CRU)	NAOI (EOF)	Max. MSLP	Min. MSLP
Western Scotland	57.25°N, 5.25°W	**0.429**	0.176	**0.413**	**−0.781**
Norway	65.25°N, 13.75°E	0.209	0.089	**0.438**	**−0.680**
Southern Spain	37.75°N, 3.75°W	0.277	0.059	**0.432**	−0.287
Central France	47.25°N, 3.25°E	**−0.538**	**−0.359**	**0.414**	**−0.595**
Czech Republic	49.75°N, 15.25°E	−0.102	−0.066	**0.653**	**−0.700**
Balkans	42.75°N, 20.25°E	−0.028	0.286	**0.477**	**−0.529**

the MSLP–precipitation correlation maps. A positive phase of the Scandinavian pattern has an anticyclone over Scandinavia and weaker cyclonic regions over Western Europe and eastern Russia, which results in lower precipitation totals over Scandinavia and higher precipitation totals over Central Europe. For example, precipitation over eastern France in October (Fig. 10) shows a positive relationship with the Scandinavian pattern with positive MSLP–precipitation correlations over Scandinavia and negative MSLP–precipitation correlations over Western Europe. In this analysis, however, only the NAOI was used to show that gridded MSLP fields yield stronger statistical relationships with precipitation than this dominant atmospheric mode's index (i.e. the NAOI).

6 CONCLUSIONS

This paper has used correlation analysis to assess the spatiotemporal variability of large-scale climatic control (MSLP) on European precipitation. The uncovered significant hydroclimatological relationships have elucidated previous knowledge of precipitation generating mechanisms in Europe by showing the detailed variations in the MSLP–precipitation relationship across Europe. The key conclusions that may be drawn from this work are:

Throughout the year, European precipitation is associated with MSLP centres of action located over different areas of the Northern Hemisphere, with the winter yielding generally larger regions of significant correlation than the summer. In winter, precipitation showed links with extensive geographical areas, reflecting the strong influence of the large-scale climatic circulation on European precipitation (especially in northwestern Europe). In northern Europe positive MSLP–precipitation correlation exists over the Azores and negative MSLP–precipitation correlation exists near Iceland, implying a positive relationship with the NAOI; in southern Europe negative MSLP–precipitation correlation exists over the Azores and positive MSLP–precipitation correlation exists near Iceland, implying a negative relationship with the NAOI. These relationships are in agreement with those of Hurrell (1995). These correlation centres relate to the Azores High and Icelandic Low pressure systems. The large spatial scale of significant climatic circulation patterns over northwest Britain and Scandinavia is thought to be caused in part by orographic effects on precipitation. Downwind from the Atlantic (east of the Scandes or of the Alps), smaller MSLP–precipitation correlation patterns occur, showing a reduced influence of westerly winds on precipitation in these regions.

In summer, European precipitation has fewer significant correlations with the large-scale climatic circulation, implying that precipitation is produced by small-scale and shorter duration weather systems. As this study used a coarse temporal (monthly) and spatial (2.5° × 2.5°) resolution of MSLP, convective systems that are a source of summer European precipitation (Berg *et al.* 2009) would not have been well captured. However, precipitation in some parts of Europe (west of Baltic Sea, northwest Europe and eastern Mediterranean) did show strong correlations with MSLP over Greenland, similar to the SNAO, a variant of the well-established NAO.

In spring and autumn, smaller MSLP patterns than in winter are associated with European precipitation, with the winter dipole shrinking and moving during spring, to reappear in September in northern Europe. MSLP–precipitation patterns reveal an extended winter season in the far north of Europe (e.g. Scandinavia) as it lingers during spring, and the Scandinavian climate pattern (Barnston and Livezey 1987) is identified in the MSLP–precipitation correlation maps over Scandinavia and Central Europe in autumn before shifting back to an NAO-like dipole in December in the British Isles and northern France.

The gridded MSLP data yield stronger empirical relationships compared to the NAOI (whether defined as a fixed dipole or from EOF analysis) because the MSLP can capture the dynamic seasonal movement of the atmospheric areas with strongest control on European precipitation. The geographical points used in the CRU and EOF NAOI definition do not always coincide with the high MSLP–precipitation correlation areas; therefore, the NAOI is a less powerful explanatory variable of European precipitation, even during winter where its links with precipitation have previously been identified. Although atmospheric indices are useful as a starting point in investigating large-scale climatic control on European precipitation variability, the results herein suggest that finer scale (i.e. gridded) data yield stronger statistical relationships and improved knowledge of the details of this important socially-relevant hydroclimatological process.

There is a gradient in the influence of North Atlantic pressure systems over Europe, as shown by larger MSLP–precipitation correlation areas in western districts compared to eastern districts (i.e. from west to east Britain, across the European Alps and from Norway to central and east Sweden). This reflects the heterogeneities of the European land mass, in particular the barrier effect of mountain chains such as the Scandes and Alps, which limit penetration of eastward-moving rain-bearing systems resulting in smaller precipitation totals in their lee (i.e. rain shadow) and weaker MSLP–precipitation correlation patterns. As the large-scale atmospheric dynamics are most active in the winter season, this phenomenon is most notable in the winter months.

The recent availability of gridded precipitation and MSLP time series has made it possible to undertake a consistent and systematic spatiotemporal analysis of the large-scale climatic control on European precipitation. The results presented herein corroborate previous research that considered atmospheric indices (e.g. Hurrell 1995), but our findings demonstrated that an index with fixed-point definition, such as the NAOI, is incapable of explaining precipitation occurrence in certain regions (i.e. Central Europe, such as the Czech Republic; Tables 1 and 2) and seasons (i.e. summer). The identified hydroclimatological relationships could be used to evaluate climate model output to determine if the location, strength and timing of these hydroclimatological connections can be reproduced accurately by models. If climate models become capable of reproducing these hydroclimatological correlation patterns at extended forecast lead times, scientific and societal benefits could result.

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