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Witnessing North Atlantic westerlies variability from ships' logbooks (1685–2008)

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Abstract A monthly index based on the persistence of the westerly winds over the English Chanel is constructed for 1685–2008 using daily data from ships' logbooks and comprehensive marine meteorological datasets. The so-called Westerly Index (WI) provides the longest instrumental record of atmospheric circulation currently available. Anomalous WI values are associated with spatially coherent climatic signals in temperature and precipitation over large areas of Europe, which are stronger for precipitation than for temperature and in winter and summer than in transitional seasons. Overall, the WI series accord with the known European climatic history, and reveal that the frequency of the westerlies in the eastern Atlantic during the twentieth century and the Late Maunder Minimum was not exceptional in the context of the last three centuries. It is shown that the WI provides additional and complementary information to the North Atlantic Oscillation (NAO) indices. The analysis of WI series during the industrial era indicates an overall good agreement with the winter and high-summer NAO, with the exception of several multidecadal periods of weakened correlation. These decoupled periods between the frequency and the intensity of the zonal flow are interpreted on the basis of several sources of non-stationarity affecting the centres of the variability of the North Atlantic and their teleconnections. Comparisons with NAO reconstructions and long instrumental indices extending back to the seventeenth century suggest that similar situations have occurred in the past, which call for caution when reconstructing the past atmospheric circulation from climatic proxies. The robustness and extension of its climatic signal, the length of the series and its instrumental nature make the WI an excellent benchmark for proxy calibration in Europe and Greenland.

Keywords Atmospheric circulation index · Climate variability and change · Early instrumental data · North Atlantic Oscillation

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1 Introduction

The strength and direction of near-surface winds have long been recognized as an essential component of the atmospheric circulation and its variability (Namias 1950; Lamb 1995). The variability of the atmospheric circulation is usually assessed by means of climatic indices, often constructed from pressure differences between two fixed stations, for example, the North Atlantic Oscillation (NAO) index (e.g., Hurrell 1995; Jones et al. 1997). These indices provide a first approximation to the main modes of atmospheric variability (e.g., Barnston and Livezey 1987) and regional circulation patterns (e.g., Philipp et al. 2007) obtained from more sophisticated methods.

While climatic indices do not provide a full description of pressure configuration, they have become powerful tools to assess climate variability over the Euro-Atlantic sector and to understand the underlying dynamics (e.g., Trigo et al. 2002; Feldstein 2003, 2007; Hurrell and Deser 2009). In consequence, many studies have tried to extend their record as far back in time as possible in order to improve our understanding of climate variability (e.g., Jones et al. 1997). However, before the nineteenth century, our knowledge of past atmospheric variability mostly relies on different climatic proxies (e.g., Glueck and Stockton 2001; Cook et al. 2002; Luterbacher et al. 1999, 2001a; Jones et al. 2009; Trouet et al. 2009). Despite their utility, indices derived from proxies suffer from a number of limitations (e.g., Luterbacher et al. 1999; Schmutz et al. 2000; Jones et al. 2001; Cook et al. 2002; Timm et al. 2004): (1) they are often constructed from a relatively small number of predictors with time-varying availability and/or poor spatial coverage; (2) their temporal resolution typically ranges from seasonal to annual or multiannual time scales; (3) the reconstruction implies calibration against observational data under the assumption of a stable relationship between the predictors and the predictand. This process suffers from different sources of uncertainty, such as those derived from non-stationary relationships between the proxy and the local climate and between the local climate and the larger-scale atmospheric circulation itself (e.g., Schmutz et al. 2000; Jung et al. 2003; Vicente-Serrano and López-Moreno 2008). The lack of stationary relationships between atmospheric circulation and European climate can be even larger in the context of the last centuries (Jacobeit et al. 2003; Casty et al. 2005; Pauling et al. 2006; Küttel et al. 2011), and could influence the reconstructions of atmospheric circulation, as the statistical models would be sensitive to the specific time periods chosen for calibration and validation (e.g., Slonosky and Yiou 2002).

An alternative method of exploring past climatic variability without using proxies is through the systematic analysis of early instrumental data over land and historical weather reports (Jones et al. 1997; Slonosky et al. 2001; Cornes et al. 2012). Early times series of instrumental-based observations usually have daily resolution, which is much higher than that of proxies, and can improve the quality of proxy-based reconstructions by providing more extended target periods for calibration and verification. Our current knowledge of past atmospheric circulation through the analysis of historical series mostly relies on instrumental data over land (e.g., Jones et al. 1997; Cornes et al. 2012). The marine meteorological observations found in old ships' logbooks provide a less exploited source of climatic information. Such data contain first-hand and well-dated daily (sometimes sub-daily) evidence on the weather that ships encountered along their route (García-Herrera et al. 2005a; Wheeler and Garcia-Herrera 2008). In particular, wind force derived from ship logbooks has been widely employed to generate long climatic series (Gallego et al. 2005; Jones and Salmon 2005; Küttel et al. 2010). Unfortunately, the use of wind force back beyond 1750 is more problematic because of the gradual evolution of the nautical vocabulary (Wheeler and Wilkinson 2005). However, wind direction information from logbooks does not suffer from such a problem and it can be considered an instrumental observation (Jackson et al. 2000; Jones and Salmon 2005; Wheeler et al. 2009) because: (1) similar to present day, it was measured with a 32-point compass, with respect to the magnetic north; (2) it does not need subjective judgments or re-scaling to modern quantitative standards; (3) for the region in question herein the magnetic variation (difference between the true north and the magnetic north) was not so large as to require significant correction.

Long and continuous series of logbook information can be compiled for specific areas with sufficient data coverage. In particular, the English Channel is a region of intense marine activity from the seventeenth century, and extant Royal Navy logbooks for ships in the area are sufficient in number to provide a near-daily wind record since as early as 1685. This is also an area of valuable climatic information for the Euro-Atlantic sector as it is located over the main belt of extratropical westerly winds. The objective of this study is to explore past atmospheric variability over the Euro-Atlantic sector as revealed by daily marine observations. Based on direct observations of the wind direction over the English Channel, we construct a monthly index of the atmospheric circulation variability over the North Atlantic from 1685 to 2008. The so-called Westerly Index (WI) measures the persistence of the westerly winds beneath the exit zone of the North Atlantic extratropical jet-stream (Wheeler et al. 2009) and it is found to be a valuable indicator of the past European climate that is not reliant on interpretations of proxy evidence.

This paper is structured as follows. Section 2 describes the data and the methods employed to construct the WI. The assets of the WI as an indicator of the North Atlantic atmospheric variability and the European climate are given in Sect. 3. Section 4 describes the evolution of the seasonal time series of the WI for the period 1685–2008 and compares it with current knowledge of the past European climate as inferred from previous studies. The relationship between the WI and several related climate indices, and the possible reasons for their discrepancies are discussed in Sect. 5. Finally, Sect. 6 summarizes the main findings.

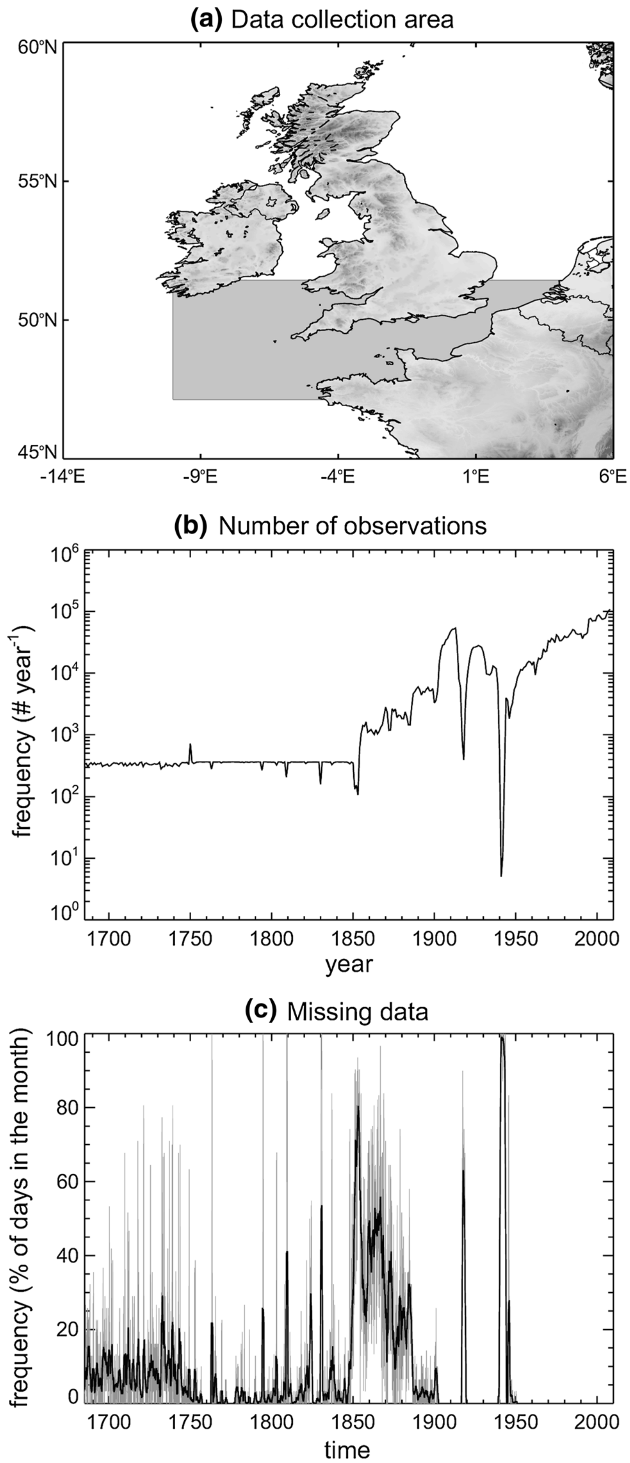


Fig. 1 **a** Area of analysis. *Grey shading* shows the selected region for the collection of wind direction observations during the period 1685–2008; **b** Total annual number of wind direction observations inside the selected region for the period 1685–2008. Note the logarithmic scale; **c** Monthly time series showing the percentage of days with missing data (*grey line*) along the 1685–2008 period. The *black line* depicts a 12-month running average

2 The Westerly Index

Royal Navy ships' logbooks from 1685 to 1850 have been searched in the archives of the National Maritime Museum at Greenwich (South East London) and the UK National Archives (Kew, Surrey). For that period, a total of 57.817 daily wind records were abstracted over the English Channel (10°W–5°E, 48°N–52°N; Fig. 1a). Fleet lists with information of vessels arranged geographically and by month allowed for the easy identification of ships in the region at any one time. Further information on the precise location of ships was given in the logbooks, since near-coastal navigation of the type common in the English Channel was made by respect to visible landmarks. The logbook search was optimized to secure at least one wind observation per day over the region. When several wind observations per day were made by a vessel, the midday (and principal) observation of each nautical day only was abstracted. Similar to modern standards, the wind direction was measured with a 32-point compass. As these data only required very minor corrections (change from Julian to Gregorian calendar before 1753 and from nautical to civil day) and magnetic variation was minimal at this time, the reader is referenced to Wheeler et al. (2009) for further information on data sources and treatment. From 1851 onwards to 2008, the CLIWOC v1.5 (García-Herrera et al. 2005b) and ICOADS v2.1 (Worley et al. 2005) datasets offered 3.392.640 observational records for the same area. They contain data that have already been processed after exhaustive analysis and interpretation of the information. Jones and Salmon (2005) have reported high consistency and homogeneity between the wind observations of CLIWOC and ICOADS datasets.

There are large differences in data availability through the analyzed period (Fig. 1b). Up to 1850, the database usually contains no more than one record per day, but, in spite of this, data were found for ∼95 % of the days in that period (Fig. 1c). From 1851 to 2008, the database provides good temporal coverage, with the exceptions of the WWI and WWII periods (see Fig. 1b), and high spatial density of observations (more than 20 observations per day are usually available from 1851 onwards). Despite the relatively high density of observations, the second half of the nineteenth century was also a period of frequent missing days (Fig. 1c).

The WI is here defined as the proportion of days per month (in %) with prevailing wind from the west (wind blowing from between 225° and 315° from true north, Wheeler et al. 2009). Having several wind observations per day, we need to define a westerly wind day in order to

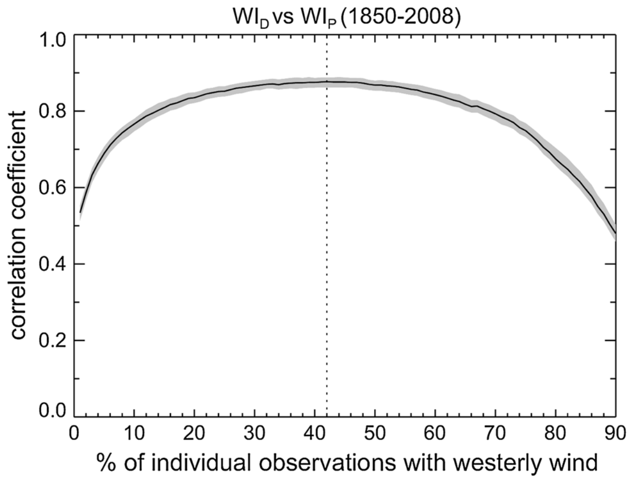


Fig. 2 Averaged correlation between 1,000 randomly degraded WI_D series computed with only one wind record per day and WI_P series computed by defining a westerly day as a day with at least a percentage *p* of wind records with westerly direction over the English Channel for the period 1850–2008. The *horizontal axis* corresponds to the value of *p*, ranging from 1 to 90 %. *Grey shading* indicates the full range of correlations obtained from the 1,000 Monte Carlo simulations. The *vertical dotted line* identifies the *p* value of maximum correlation

compute the WI. When only one daily observation was available, the day was classified from the wind direction of that single observation. If several observations per day were available, a westerly wind day was said to occur when a minimum percentage *p* of the total number of observations of that day corresponded to westerly wind. The specific value of *p* was derived from a process aiming at optimizing the homogeneity of the series between periods with a single daily observation and periods with multiple observations. Accordingly, different provisional WI series (WI_P) were computed, each one derived from a *p* value between 1 and 90 % for 1850–2008, since it was a period with a high frequency of multiple daily observations. Furthermore, we derived 1,000 "degraded" WI series (WI_D) for the same period, all constructed from only one observation per day. Each of the WI_D series is expected to be different because the single observation is randomly selected among all the wind observations available each day.

The WI_P series were correlated with the 1,000 WI_D series and the mean correlation as a function of *p* is displayed in Fig. 2. The correlation increases monotonically reaching a maximum for the value of $p = 42$ % (r = 0.88; $p < 0.01$) and then, it decreases steadily. Similar results were found for different periods (1750–2008, 1800–2008, 1900–2008), thus suggesting that this optimum *p* value is not sensitive to the frequency of daily observations within the English Channel. As a consequence, the final WI series has been constructed by considering a westerly wind day that with at least 42 % of the daily wind direction observations corresponding to westerly wind. Note, however, that all *p* thresholds led to similar and significant correlations ($p < 0.01$, grey shading in Fig. 2). This relative lack of sensitivity to the exact value of *p* confirms a spatially coherent pattern of wind direction within the English Channel, further supporting the selection of that area to characterize atmospheric variability over the North Atlantic.

Finally, daily WI values were aggregated into the monthly statistics. The monthly WI values are based on the proportion of westerly days with respect to the total number of observations for each month. Months with more than 10 days (~33 % of the month) of no wind direction observation were labeled as missing, although different thresholds did not change the conclusions drawn herein.

To identify temporal inhomogeneities in the monthly WI series, an analysis of structural changes on the mean was performed based on the Chow test (Chow 1960) by assessing deviations from stability in running linear regressions. Potential breakpoints are identified when the value of the coefficient of the regression shifts from one stable regression relationship to a different one and the null hypothesis of no structural change is rejected based on an *F* statistic (Zeileis et al. 2003). For the WI, the test identified a potential breakpoint in October 1762. However, its magnitude was quite small (a jump on the mean WI of less than 2 %, equivalent to 0.62 days of westerly wind in a month) and it was not coincident with any period of changes in the data sources or in the density of observations. Additionally, to assess potential inhomogeneities in the variance of the WI, a wavelet variance analysis was carried out by applying the Maximal Overlap Discrete Wavelet Transform (e.g., Percival 2008), which allows the decomposition of the WI time series into additive subseries, from high-frequency (intermonthly) to low-frequency (interdecadal) signals. The results revealed that high-frequency variability was typically below 15 % and fairly stable in time, without sudden changes in variance. The above results suggest negligible effects arising from temporal inhomogenities in the availability and spatial coverage of wind observations.

Finally, the WI series was compared with daily data from the twentieth century reanalysis (20CR, Compo et al. 2011) for the period 1871–2008. By using regional mean values of the zonal and meridional wind components from the 20CR over the English Channel on a daily basis, we derived a suite of 20CR-based monthly time series of WI (WI_{20CR}) at several vertical levels. The best agreement between the WI and all WI_{20CR} series was found for 10-m, the lowest vertical level with wind data in the 20CR. The WI and the 10-m WI_{20CR} series displayed a smoothed seasonal cycle on the frequency of westerlies, with values around 35 % and standard deviations of 17 % for most of the year. It is worth noticing that, unlike the zonal wind component and the wind speed, maximum WI values are found in summer, in response to the seasonal March in intensity and location of the Iceland low and the Azores high (Wheeler et al. 2009). Maximum differences between the climatological monthly values of WI and WI_{20CR} amount to 5 % ($\sim$1–2 westerly wind days in a month) and their temporal correlation for the 1871–2008 period remains very high throughout the whole year. This comparison does not fully allow for testing the reliability of the WI, since the 20CR also relies on ICOADS pressure data and hence they are not fully independent in the sense that the wind observations employed to construct the WI for 1871–2008 and the 20CR-derived wind data have one common platform within their data sources. However, the good agreement between WI and WI_{20CR} points to negligible influences of local factors and a strong spatial coherence among the directional wind fluctuations over the English Channel (recall that the WI_{20CR} was obtained from regional mean values while the WI relies on individual observations). This result suggests that the WI can be considered a robust measure of near-surface wind direction over the area-averaged English Channel. The derived product provides an almost continuous monthly series for between 1685 and 2008 and represents the longest currently available instrumental climate series of atmospheric circulation.

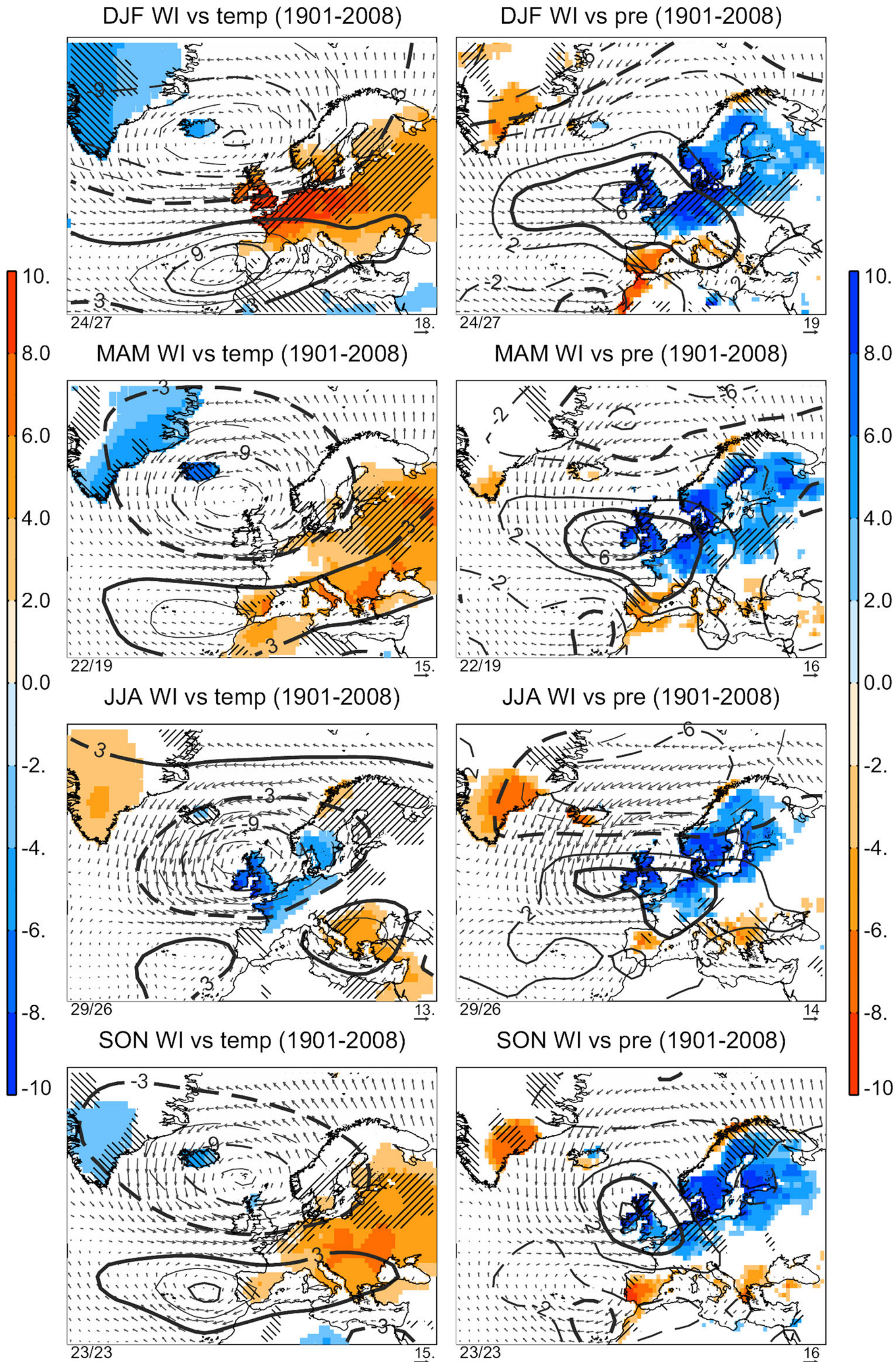
3 The WI as an indicator of the European climate

The relation of the WI with the climate of the North Atlantic has been examined using monthly precipitation totals from the GPCC (Rudolf and Schneider 2005), and monthly mean near-surface temperature from the CRU TS3 (Harris et al. 2012) for their common period of data availability (1901–2008). Both datasets are confined to land-based observations. Furthermore, daily mean fields of geopotential height, temperature, humidity and zonal and meridional wind fields at different pressure levels have been obtained from the 20CR in order to compute the following variables: (1) the vertically integrated moisture transport through the 1,000–500 hPa layer; (2) the moisture convergence between 1,000 and 500 hPa; (3) temperature advection at different levels and; (4) the storm-tracks, defined as the standard deviation of the high-pass filtered (2–7 days) geopotential height at 500 hPa. This suite of temperature—(Fig. 3, left panels) and precipitation—(Fig. 3, right panels) related fields has been composited for high (low) phases of the WI defined as those cases when the normalized seasonal WI value is above (below) 0.75 (−0.75) standard deviations. This definition provides a balanced and sufficient number of cases in each composite. Seasonal scaled composites were computed by scaling the mean by the squared root of the sample size and dividing by the standard deviation of the composited sample. The variables so-composited are dimensionless but they provide a more robust signal than traditional composites, as they are less sensitive to outliers (Brown and Hall 1999; Küttel et al. 2011). The significance of the scaled composite difference between high and low WI values was tested against climatology by using a Monte Carlo test of 1,000 realizations, in which two subsets with a number of cases equal to those of high and low WI phases were randomly chosen from among the available period.

Figure 3 shows the seasonal scaled composite differences of temperature- (left panels) and precipitation- (right panels) related variables for high minus low WI phases. WI fluctuations are associated with a meridional dipole in the anomaly pressure field (Fig. 3, left panels, contours), indicating lower-than-normal pressures dominating over the eastern North Atlantic and positive pressure anomalies in subtropics during positive WI phases (enhanced frequency of westerlies), and an opposite circulation pattern during negative WI phases (reduced frequency of westerlies). This dipolar signature is observed throughout the whole year, but it changes in intensity, extension and location depending on the season.

Figure 3 (left panels) shows that for a high phase of WI, the increased frequency of westerlies is responsible for an enhanced advection of warm air during winter (hatched areas), producing warmer conditions over a latitudinal band that extends from the northern Mediterranean to northern Russia (shaded areas). A large area of strong negative anomalies develops over Greenland as a result of the consequent advection of cold air from polar latitudes. Transitional (spring and autumn) seasons have WI-related temperature anomalies of the same sign as those observed in winter, although smaller in magnitude and spatial extension. In summer, the reduced intensity of the zonal wind is mirrored in the spatial extension of the circulation anomaly pattern and the associated temperature field. A high frequency of summer westerlies is associated with reduced temperatures over northwestern Europe, in agreement with cyclonic conditions over the UK and enhanced cloudiness that reduces shortwave radiation.

The changes in precipitation-related variables during opposite phases of the WI reveal a tripolar structure in precipitation for the whole year (Fig. 3, right panels, shading areas). WI-related precipitation anomalies are accompanied by changes in the level of synoptic activity with high (low) WI values reflecting an enhanced (weakened) synoptic eddy activity over the eastern part of the North Atlantic and northern Europe (contour lines). These anomalies are reflected in pronounced changes in the



transport and convergence of atmospheric moisture (hatched areas) and therefore, precipitation. Thus, during phases of high WI, enhanced precipitation extends from northwestern Europe to northern Russia and much of the Scandinavian Peninsula, whereas precipitation deficits span the northern Mediterranean and southeastern Greenland, with opposite patterns occurring during low WI phases. These results indicate that the WI fluctuations can be interpreted in terms of temperature and precipitation changes over large areas of Europe and hence the long record of the WI is expected to be a valuable indicator of past European climate for the last three centuries.

◀**Fig. 3** Seasonal differences between scaled anomaly composites of high and low WI phases for: *Left panels* geopotential height at 500 hPa (Z500, contours), land surface temperature (*shaded areas*), 500 hPa wind (*arrows*) and 500 hPa temperature advection (*hatched areas*); *Right panels* storm tracks (2–5 high pass filtered Z500 variance, contours), land precipitation (*shaded areas*), 1,000–500 hPa vertically integrated moisture transport (*arrows*) and 1,000–500 hPa moisture convergence (*hatched areas*). All units are dimensionless. *Solid (dashed) contour lines* represent positive (negative) values, with thick contours indicating significant differences from climatology at $p < 0.05$. Only temperature and precipitation differences that are significant at $p < 0.05$ are shown and gridpoints with climatological mean precipitation below 10 mm are omitted. Cross-hatched areas with lines orientated 45°/−45° from the east indicate (*left panels*) warm/cold temperature advections and (*right panels*) divergence/convergence of moisture fluxes with significant differences at $p < 0.05$. The length and size of the arrows is proportional to the magnitude of the anomaly of (*left panels*) wind at 500 hPa and (*right panels*) moisture flux. A reference value is shown in the *bottom right corner* of each panel. For better readability of the figure, *hatched fields* (*arrows*) are only displayed over land (ocean). Numbers in the *left bottom corner* of each panel represent the number of cases employed in the composite of high/low WI. Significance is assessed with a 1,000-trial Monte Carlo test

4 The WI as a record of the past European climate

In this section, we look at the WI record in the context of the history of the European climate found in the literature. Figure 4 shows the time series of the standardized seasonal mean WI anomalies. Cumulative normalized seasonal anomalies are also shown (black thick lines) to better identify periods of predominant positive (negative) anomalies, which are evidenced by increasing (decreasing) phases.

The beginning of the WI series is coincident with the Late Maunder Minimum (LMM, ca. 1685–1715), a reported cold period over large areas of Europe, particularly during winter and spring (Wanner et al. 1995; Luterbacher et al. 2001b, 2004; Xoplaki et al. 2005). The winter WI shows for this period predominantly negative but not particularly extreme values, suggesting that the LMM was not characterized by a persistent low frequency of westerlies in the English Channel. This is compatible with the idea that the LMM exhibited recurrent winter anticyclonic conditions over eastern Europe (Luterbacher et al. 2001b; Jacobeit et al. 2003; Pauling et al. 2006), which would not have a dramatic impact on the circulation over the English Channel. The unfiltered WI values also confirm that the late seventeenth century was characterized by great variations in zonal flow over Europe. For example, the winter of 1696–1697 and the summer of 1698 were associated with extreme cold conditions, including harvest failures and famines in England and Scandinavia, while 1698–1699 was a warm winter (Slonosky et al. 2001). This is captured by the value of the normalized seasonal anomalies of WI ($WI_{DJF}(1697) = -0.98$, $WI_{JJA}(1698) = 0.85$ and $WI_{DJF}(1699) = 1.46$, respectively).

The suggested shift towards an increased European rainfall in northern Europe after the LMM (Luterbacher et al. 2001b; Jacobeit et al. 2003; Pauling et al. 2006) is also confirmed by the WI, which shows preferred positive values up to the mid-eighteen century. The second half of the eighteenth century marks a turning point toward more frequent negative phases of the WI, in accordance with the reduced zonality during the 1780s and 1790s reported by Slonosky et al. (2000). This is also evidenced from a trend analysis as a statistically significant decreasing trend ($p < 0.05$) of the WI during the second half of the eighteenth century (not shown). Interestingly, one of the coldest winters recorded in the Central England Temperature series (1783–1784) occurred during this period (Manley 1974). Recently, this episode has been attributed to a combination of a negative phase of the NAO and an El Niño event (D'Arrigo et al. 2011), rather than to the Laki eruption, as previously thought. The strong negative value of the normalized WI for that winter (−1.10) supports the more recent interpretation. On the other hand, the WI shows strong variability during the eighteenth century and relatively smaller fluctuations since the nineteenth century. Previous studies based on weather reports (e.g., Kington 1980) and long-instrumental series (e.g., Slonosky et al. 2000; Cornes et al. 2012) have also suggested that the variability of the atmospheric circulation was greater before than after the early-to-mid nineteenth century.

A period of predominantly low frequency of westerly flow through the English Channel marks the turn to the nineteenth century, and is particularly pronounced between the 1800s and the 1830s, persisting until the second half of the nineteenth century. Again, this tendency supports the hypothesis of periods of weak zonal circulation (Slonosky et al. 2000) and frequent North Atlantic blocking (Moses et al. 1987). Except for summer, the prevailing negative WI phases are in agreement with a concurrent temperature drop reported by Dobrovolný et al. (2010) over central Europe. The WI also captures known extreme episodes such as the cold winter of 1837–1838 (WI = −1.84) and the extremely warm winter of 1833–1834 (WI = 2.18).

The analysis of the WI record also reveals some contradictions with previous studies for the first half of the nineteenth century. For example, an increase of the summer soil moisture conditions over large parts of Europe has been documented for these years (Briffa et al. 2009). While this situation should be expected to correspond to enhanced precipitation and a high frequency of westerly winds (Fig. 3, right panels), the summer WI reveals no positive values over this period. As a possible explanation, it has been suggested that the period around 1800 displayed a breakdown of the relationship between circulation patterns and the associated climatic anomalies and thus, that climatic boundary conditions (rather than those arising from circulation changes) may have played a dominant role at that time (Jacobeit et al. 2003). The absence of a marked tendency to positive WI values over this period (Fig. 4) lends support to the latter hypothesis.

The years around 1850 have been identified as a period of profound circulation changes (Manley 1974; Jacobeit et al. 2001; Slonosky et al. 2000; Beck et al. 2007; Küttel et al. 2011) and taken thereby as the transition period from the Little Ice Age to the recent warmer climate. Frequent transitions in winter quasi-stationary weather regimes (Luterbacher et al. 1999; Casty et al. 2005; Philipp et al. 2007), and an enhanced variability of European mean climate parameters (e.g., Pauling et al. 2006; Casty et al. 2007) have been also reported for that time. This period is characterized by slowly increasing values of the winter WI, particularly during the second half of the nineteenth century. By the first third of the twentieth century all (but autumn) seasonal series of WI show an increase in the frequency of westerliness, followed by a decline of the WI up to 1950s–1970s and a moderate recovery thereafter. The last decades of the twentieth century, and in particular the 1980s and 1990s arise as a period of relatively higher frequency of westerlies than other periods of comparable length in the 1685–2008 record, but these decades were not exceptional in the context of the last centuries, a result already found from regional indices of atmospheric circulation (e.g., Jones et al. 1997; Cook et al. 2002; Luterbacher et al. 2001a; Slonosky and Yiou 2002; Jacobeit et al. 2003; Cornes et al. 2012). However, and different to other climatic indices, the WI does not display a clear shift to persistent positive values during the 1970s–2000s indicating that, in the context of the last three centuries, these recent decades were not particularly anomalous in the frequency of westerlies over the English Channel.

5 The WI and the NAO

From the previous sections, we note that the WI-related patterns shown in Fig. 3 resemble those obtained for the NAO (e.g., Trigo et al. 2002). There are also some similarities to the historical evolution of the WI (Fig. 4) and those of the NAO indices (e.g., Luterbacher et al. 2001a). Thus, for example, in winter, the anomaly dipole associated

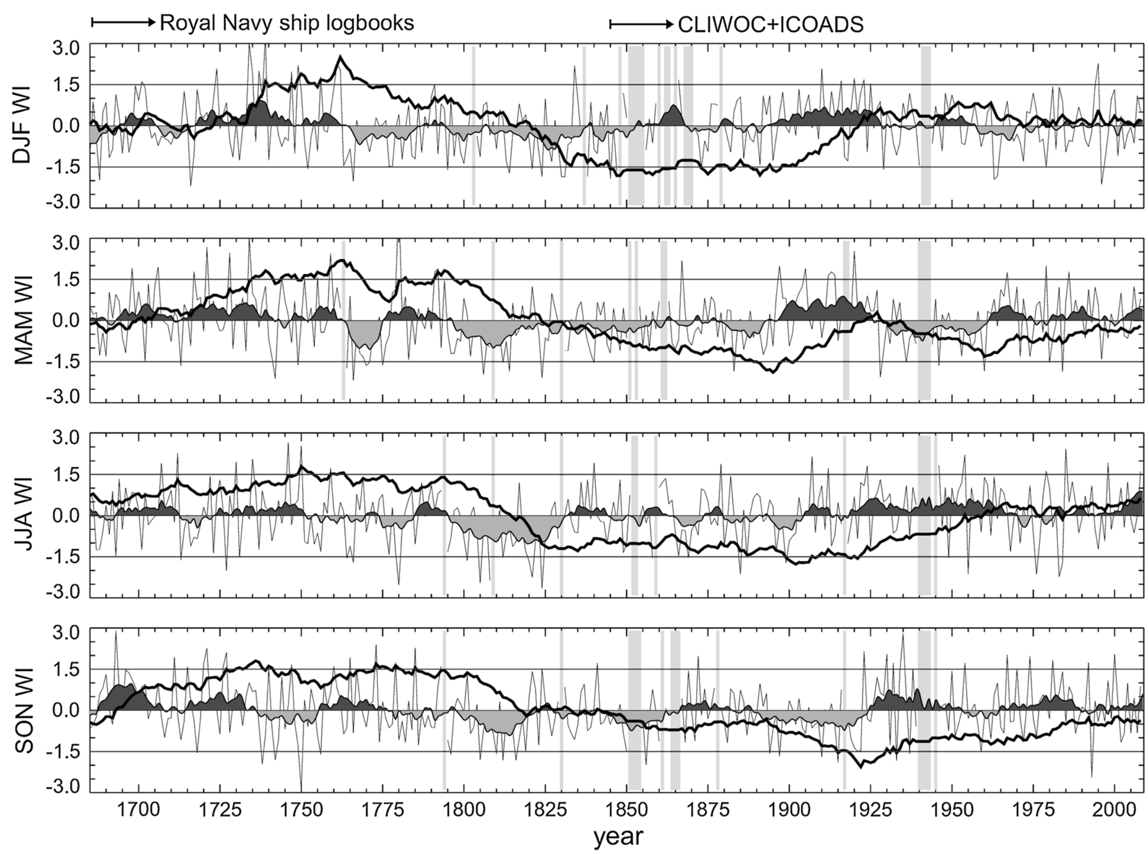


Fig. 4 Standardized seasonal series of WI for 1685–2008 (*grey line*) and the 11-year running average superimposed, with *dark/light grey shading* highlighting periods above/below the 1685–2008 average. *Horizontal lines* indicate ±1.5 standard deviations relative to the 1685–2008 period. *Black thick line* shows the cumulative normalized seasonal anomalies. *Vertical gray bars* indicate periods of missing data (i.e. seasons with at least one missing month)

with WI anomalies (Fig. 3, left panels, contours) spreads through most of the Atlantic in an NAO-like fashion. Moreover, the anomalous anticyclonic circulation associated with negative phases of the summer WI is north of UK and well co-located with the southern lobe of the summer NAO (Folland et al. 2009). This suggests that the WI shares some basic features with the winter and summer NAO and seems to be able to capture the seasonal changes in the NAO pattern (Portis et al. 2001).

However, there are also conceptual differences between the NAO and the WI. These arise from fundamental differences behind the concepts of wind force and the persistence of wind direction. Indices based on meridional pressure differences measure large-scale pressure gradients, and hence the strength of the geostrophic wind along a given direction, but they contain less information on the persistence of the wind direction. On the other hand, the WI is constructed from wind direction measures only, lacking any wind force information. Thus, discrepancies can arise between the WI and the NAO, particularly when the meridional wind component deviates appreciably from its climatological mean. On the other hand, the WI is based on real wind observations and hence, it includes an ageostrophic component of the real wind that is missed by indices based on latitudinal pressure gradients. Additional discrepancies are expected between the WI and the NAO because: (1) the WI provides a measure of the prevailing monthly conditions, while the NAO is often based on the monthly mean; (2) the NAO provides a measure of the zonal wind over areas larger than the English Channel. Therefore, one should not expect a perfect correspondence between the WI and circulation indices, such as the NAO, that provide some measure of the geostrophic zonal wind.

In the following subsections, the relationship between WI and several indices based on latitudinal pressure gradients will be addressed in more detail, paying special attention to the NAO. This comparison exercise does not aim to evaluate the reliability of other indices, nor verifying the WI using other climatic indices. Instead, it shows that the WI and other climatic indices are complementary in our understanding of the past atmospheric variability. The discussion will focus on the differences between: (1) the spatial patterns of temperature and precipitation associated with the WI and the NAO and; (2) the time series of the directional index (WI) and different climatic indicators of the strength of the westerly flow.

5.1 Spatial signatures

We start assessing differences in the spatial patterns of temperature and precipitation anomalies associated with the NAO and the WI. Because of its importance for climate reconstructions, the stationarity of these linkages is also evaluated by performing running 30-year correlations between these indices and precipitation and near-surface temperature over land through the twentieth century. For all seasons, except summer, the seasonal NAO index was obtained from the monthly NAO index of Jones et al. (1997). For the summer season, we used the high-summer (July–August) NAO index defined by Folland et al. (2009), since it is fundamentally different from the canonical winter NAO pattern. The summer NAO is the main mode of atmospheric variability over the Euro-Atlantic sector during the high-summer and it is placed further north and displays a smaller spatial scale than its winter counterpart (Folland et al. 2009). Such a dominant pattern is not observed during June (e.g., Bladé et al. 2012) and hence, comparisons between the summer NAO and the summer WI will hereafter use the summer NAO index of Folland et al. (2009) and will be confined to the high-summer (JA) season.

Figure 5 displays the percentage number of running 30-year intervals within the 1901–2008 period with significant correlations ($p < 0.01$) between each seasonal index and near-surface temperature and precipitation. For both indices, there are regions that only display temporary connections with temperature and/or precipitation through the twentieth century. This could reflect non-stationary relationships of the atmospheric circulation with the local climate. Previous studies have already reported that the occurrence of a particular canonical circulation pattern can be related to a wide range of flow configurations (Jacobeit et al. 2003; Philipp et al. 2007; Beck et al. 2007; Küttel et al. 2011). These internal (also called within-type) changes can arise from dynamical causes (changes in strength, location or spatial extension of the centres of action) that lead to variations in the circulation-climate relationship. Climatic factors, that is, the range of precipitation and temperature values that are compatible with a given circulation pattern, are an additional source of lack of stationarity (e.g., temperature advection that largely determines the temperature anomalies depends on the background thermal gradients and on sea surface temperatures, see Küttel et al. 2011).

Despite the variety of circulation patterns that can be associated with anomalous westerlies over mid-latitudes, Fig. 5 suggests that the WI signatures, particularly those of precipitation, have been stationary throughout the twentieth century, mainly over northwestern Europe and the British Islands. The impacts associated with WI fluctuations are more stationary during winter and summer than in transitional seasons, but in the case of precipitation, robust links are observed all year round and they do not seem to be as sensitive to the specific period as in the case of temperature. The opposite seems to occur for the NAO, which shows more widespread and stable correlations with

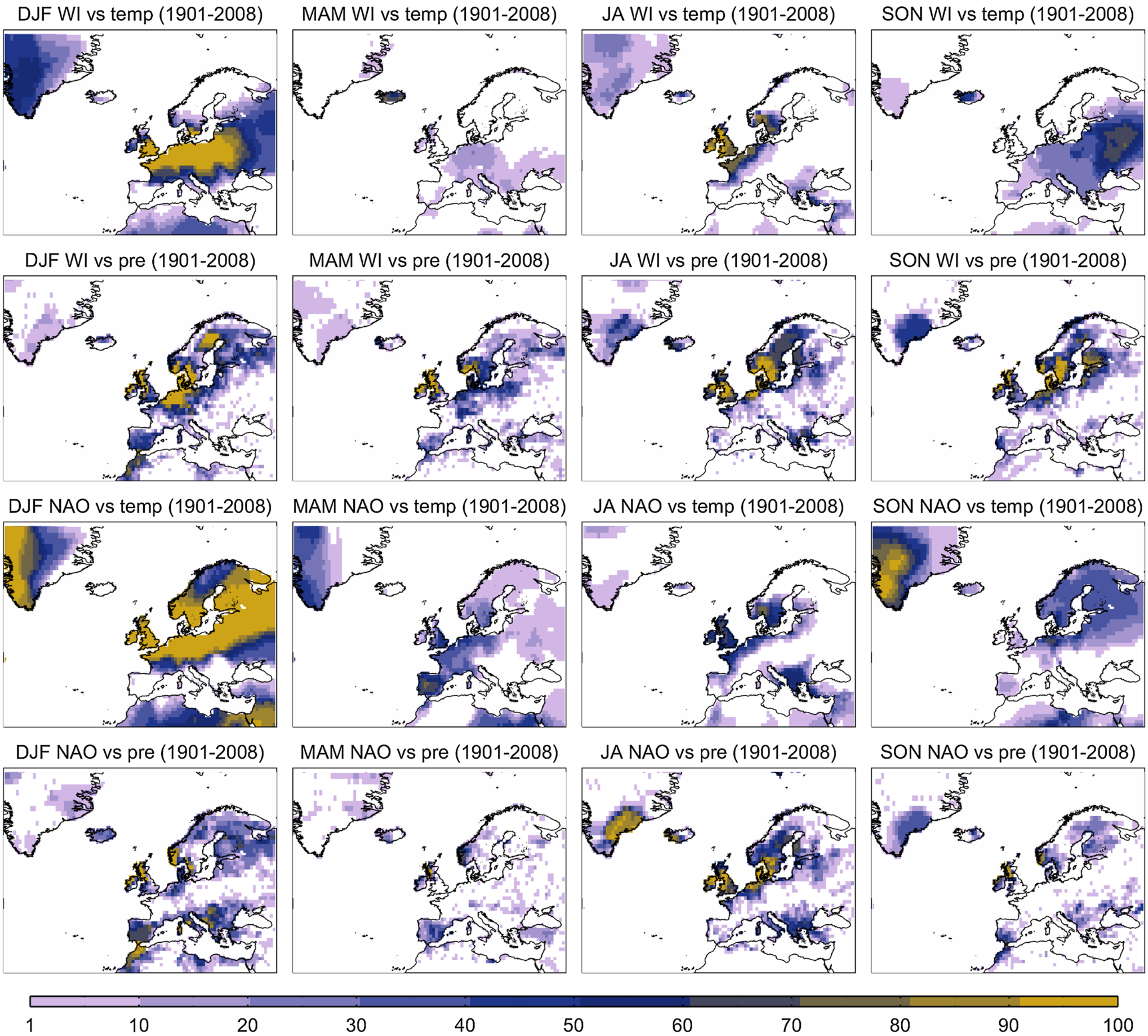


Fig. 5 Percentage number of running 30-year intervals within the 1901–2008 period with significant correlations ($p < 0.01$) between seasonal: (*the first two rows of panels*) WI; (*the last two rows of panels*) NAO and land-based observations of near-surface temperature (*odd rows*) and precipitation (*even rows*). The percentage is computed with respect to the total number of 30-year periods with complete data. Pearson and Spearman correlations are computed for temperature and precipitation, respectively. Correlation coefficients with temperature take into account the reduction in degrees of freedom due to 1-lag series autocorrelation. Seasonal NAO indices are derived from Jones et al. (1997), except the summer NAO, which uses the Folland et al. (2009) high-summer (JA) NAO index. See text for details

temperature than with precipitation (except over Iberia in winter), particularly during transitional seasons. Thus, while the WI signature in precipitation is robust through the year, this is not the case for NAO, pointing to differences between the impacts associated with NAO and WI. These results suggest that directional circulation indices such as the WI could be valuable indicators of north-central European precipitation. The lack of wind force information on the WI is most likely responsible for the weaker signals in temperature, since temperature anomalies are largely determined by advection. In this context, WI and NAO arise as complementary circulation indices, since they seem to share some commonalities, but also differences, as is evident from their associated impacts. This is in agreement with Slonosky et al. (2000) and Slonosky and Yiou (2002), who argued that the NAO was not a perfect measure of the zonality of the flow, particularly in other seasons than winter. Therefore, their combined information, depending on the region and variable considered, could be exploited to better understand climatic anomalies in the past.

5.2 Time series analysis

To explore further the added value of the WI, we compared it with winter and summer series of the NAO. Since purely instrumental NAO indices do not extend back beyond the first half of the nineteenth century (eventually the end of the eighteenth century, Jones et al. 1999), we also employed other regional indicators of the zonal wind over the North Atlantic. They include different (sometimes proxy-based) versions of the NAO and the Paris-London index, which measures the instrumental pressure difference between these two locations and is strongly linked to the NAO (Cornes et al. 2012). Table 1 lists the climatic indices along with its definition and data source. Purely proxy-based indices are subject to the uncertainties described in the introduction, which makes it difficult to interpret the causes behind the periods of low agreement with the WI. In particular, uncertainties introduced by the limited number and quality of predictors and by non-stationarities have been considered fundamental for the gradual loss of correlation between different NAO reconstructions before the instrumental period (Schmutz et al. 2000; Cook et al. 2002; Luterbacher et al. 2001a). Instrumental-based indices do not suffer from such problems, but in the case of indices constructed from historical instrumental data, there are several sources of uncertainty that could affect their reliability before approximately 1850. They include changes in the type or emplacement of the barometer or errors in the contemporary temperature records, which are required to correct the pressure readings (for further details see the references in Table 1). While these are potential sources of discrepancy with the WI that need to be taken into account in the comparison, our objective is not to identify potential periods of questionably reliability in other indices, but instead, to highlight the differences between the WI and other complementary climatic indices.

To avoid misleading conclusions arising from the aforementioned shortcomings of proxy-based climatic indices, we start by performing an intercomparison of the WI with purely instrumental indices. To do so, running correlations for windows of different length spanning between decadal and secular time scales were computed (Fig. 6). For winter, both the NAO index of Jones et al. (1997) for 1821–2008 (Fig. 6a) and the Paris-London index of Cornes et al. (2012) for 1692–2007 (Fig. 6b) exhibit similar patterns and significant correlations with WI over long time periods, with the common exception of 1855–1895. A similar result (not shown) is obtained with the revisited Iceland-Iberia NAO index of Vinther et al. (2003) for 1821–1999. The Paris-London index extends

Table 1 Climatic indicators of the strength of the westerly flow over the North Atlantic

Index	Resolution and period	Definition	Data sources	References
NAO	Monthly (1821-present)	Gibraltar-Iceland	Early instrumental pressure	Jones et al. (1997)
NAO	Monthly (1821–1999)	Iberia-Iceland	Early instrumental pressure (Cádiz and San Fernando observations for 1821–1856)	Vinther et al. (2003)
Paris-London	Monthly (1692–2007)	Paris-London	Early instrumental pressure	Cornes et al. (2012)
NAO	July–August (1850–2009)	1st EOF of SLP	European-North Atlantic mean SLP dataset (EMSLP) by Ansell et al. (2006)	Folland et al. (2009)
NAO	Monthly (1658–2001) and Seasonal (DJF) (1500–1658)	PCR with the Azores-Iceland NAO	Instrumental station pressure, temperature and precipitation (1658-on) and documentary proxy data. Only instrumental data since 1900	Luterbacher et al. (2001a)
NAO	Winter (DJF) (1430–1984)	MR with the Lisbon-Iceland NAO	Tree-rings from Finland and Morocco and Greenland Ice Sheet Project 2 (GISP2) $\delta^{18}O$ and snow accumulation records	Glueck and Stockton (2001)
NAO	Winter (DJFM) (1400–1979 and 1980–2001 instrumental)	PCR with the Gibraltar-Iceland NAO	Tree-rings from eastern North America, Europe and Morocco and Greenland ice cores	Cook et al. (2002)
NAO	July–August (1706–1976)	PCR with the summer NAO	Tree-ring chronologies in Norway and the UK	Folland et al. (2009)

Columns indicate the temporal resolution and period of the time series, the definition of the index, the data sources and the original reference, respectively. The first four indices are based on instrumental data, while the rest are derived from proxies. The third column identifies the method used to obtain the indices, including the standardized pressure difference between sea level pressure (SLP) at two locations or the first empirical orthogonal function (EOF) of regional SLP anomalies for the instrumental case, and multiple regression (MR) or principal component regression (PCR) of proxies (predictors) onto an instrumental-based NAO index (predictand) for the reconstructed series further back to the late seventeenth century, but the high frequency of missing data prevents meaningful correlations before 1750. Therefore, we also show the multiproxy 1500–2001 NAO reconstruction of Luterbacher et al. (2001a) because it includes as predictors long instrumental series (and instrumental-only data since 1900, Fig. 6c). The winter correlation pattern of this multiproxy NAO index is similar to those of the NAO and the Paris-London instrumental-only indices over their respective overlapping periods, but with slightly lower correlations. This agreement was expected because the multiproxy NAO reconstruction includes the pressure series of Paris (since 1671), London (since 1697), and Gibraltar-Reykjavik (since 1821). Interestingly, Fig. 6c shows two additional multidecadal periods over which the correlation between NAO and WI breaks down: 1700–1730 and 1760–1790. The latter is also a multidecadal interval with weak (but still significant) correlations between WI and the Paris-London index.

For summer, the instrumental NAO index of Folland et al. (2009) shows significant negative correlations with the WI for all time scales above 30-years during their common record (1850–2008, Fig. 6d). The negative correlation is due to the definition of the summer NAO (Folland et al. 2009), which differs from the traditional summer NAO index used by Jones et al. (1997). In fact, the summer NAO index of Jones et al. (1997) displays only marginal correlations with WI for summer (not shown), as does the summer NAO of Luterbacher et al. (2001a) (Fig. 6f). A relatively good degree of coherence is also found between WI and the Paris-London index during summer, but restricted to the period from 1850 (Fig. 6e). A feasible explanation for the discrepancy before 1850 is that early pressure data for summer suffer from larger potential errors than in other seasons because the temperature measurements required for correcting the barometer readings were not always recorded near the barometer, which would affect the reliability of the summer Paris-London index

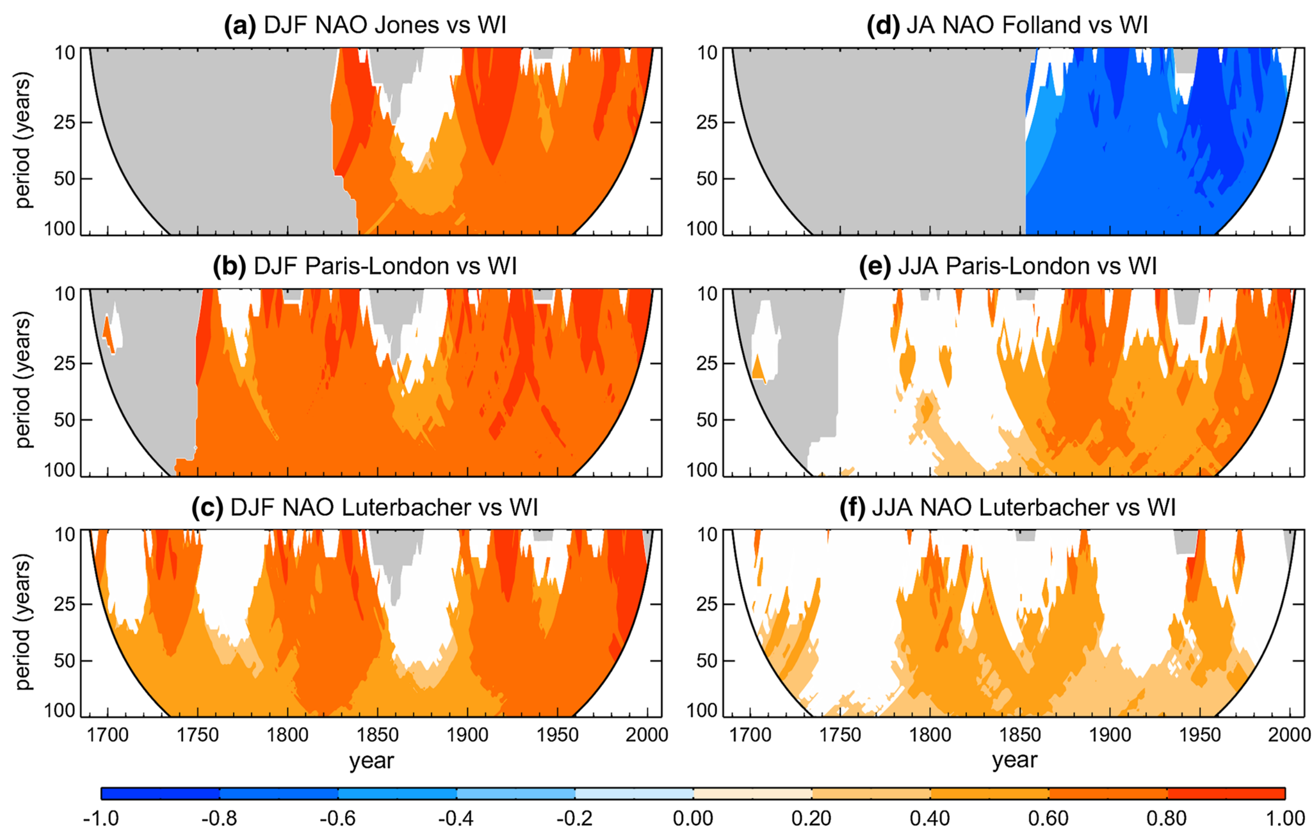


Fig. 6 Running Pearson's correlation coefficient for variable window width (*y-axis*) within the 1685–2008 period (*x-axis*) between the WI and different climatic indices of the North Atlantic circulation, including: **a** the winter NAO index of Jones et al. (1997); **b** and **e** the winter and summer Paris-London index of Cornes et al. (2012); **c** and **f** the winter and summer reconstructed NAO indices of Luterbacher et al. (2001a); **d** the instrumental high-summer NAO index of Folland et al. (2009). The *vertical axis* indicates the number of years over which the running correlation is computed. The *horizontal axis* indicates the centre of the window used to compute the running correlation. Only statistically significant correlations ($p < 0.05$) are represented. *White areas* show non-significant correlations at $p < 0.05$. Correlations are only computed when the sample size of available (not missing) data is equal or larger than 10 and it exceeds the half size of the window. *Grey shading* indicates periods of missing correlations (i.e. those that do not satisfy the previous criterion). Correlation coefficients take into account the reduction in degrees of freedom due to 1-lag series autocorrelation

(Cornes et al. 2012). The interesting results obtained from this cross-comparison stress that, overall, the WI is able to capture fluctuations of the major mode of atmospheric circulation variability over the Euro-Atlantic sector during both winter and summer despite the substantial structural differences in their spatial patterns. However, the aforementioned periods of discrepancy stress again differences between the directional WI and climatic indicators of the strength of the westerly wind, which require further analysis. This will be the subject of the next subsection.

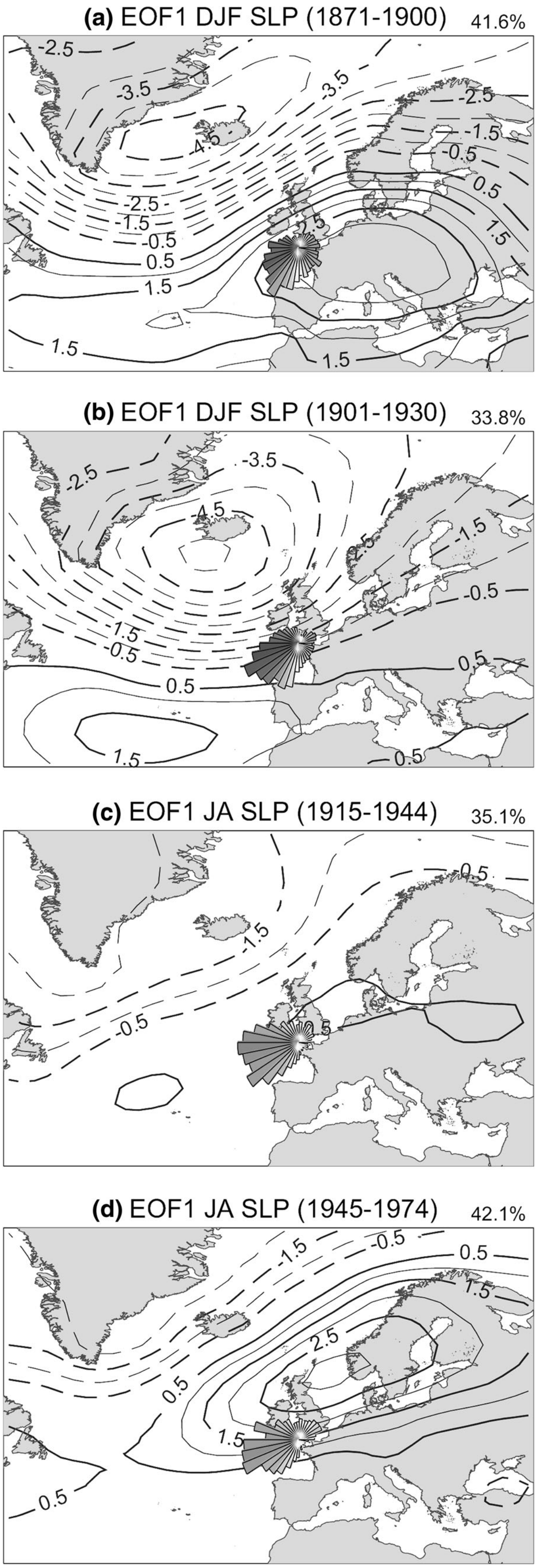
Concerning proxy NAO reconstructions, only that of Luterbacher et al. (2001a) for winter (Fig. 6c) shows an overall good agreement with the WI. In general, other proxy seasonal NAO reconstructions such as those of Cook et al. (2002) (1400–1979, with instrumental data appended for 1980–2001) or Folland et al. (2009) (1706–1976) correlate poorly with the WI, or reveal no correlation at all (Glueck and Stockton 2001, 1430–1984 period, figures not shown). Interestingly, for those reconstructed NAO series that were updated with instrumental data, the correlation with the WI generally improves during that instrumental period. In contrast, purely proxy NAO reconstructions extending into the present display significantly lower correlations with WI than their instrumental counterparts.

5.3 Discussion on the connection between WI and NAO

It should be noted that even the instrumental NAO indices show periods of weak correlation with the WI. It is also notable that the phases of non-significant correlations are not randomly distributed but aggregated in discrete periods common to all series. This could indicate statistical sampling errors, but these decoupling periods occur independently of the sources from which the WI is constructed or the number of observations available. Alternatively, the weak correlations may indicate that different indices have not always been connected to the westerly circulation in its current form. This would involve periods of non-stationarity in the connection circulation-climate and hence that proxy-based NAO reconstructions could be biased for specific periods of the preinstrumental times (Cullen et al. 2001; Schmutz et al. 2000; Timm et al. 2004). To explore possible reasons for the reported decoupling, we selected the 30-year periods of maximum and minimum correlation between WI and the instrumental NAO indices of Jones et al. (1997) and Folland et al. (2009) for winter and summer respectively within the 1871–2008 period covered by the 20CR. To identify the dominant pattern of SLP fluctuations for each case, we performed a principal component analysis (PCA) for the winter and summer seasons separately, and represented the SLP fields regressed onto the first EOF. Following the standard definition of the summer NAO (Folland et al. 2009), the PCA analysis has been applied to July–August mean SLP anomalies (computed with reference to the corresponding 30-year climatology), while the winter NAO is defined for the standard winter (December-to-February) season. The spatial area employed for the PCA analysis also differs with the season. Thus, the region (90°W–40°E, 20°N–70°N) is used for the winter NAO (as in Hurrell and Deser 2009), while the (90°W–30°E, 40°N–70°N) region is employed for the summer NAO (following Bladé et al. 2012). Small changes in the temporal or spatial domain did not change the conclusions. Additionally, for each 30-year period, we have computed the wind rose from daily values of regional mean zonal and meridional wind speed over the English Channel using 20CR data.

Figure 7a, b show the SLP patterns of the first EOF and the wind rose corresponding to the poorest (1871–1900) and highest (1901–1930) correlated 30-year winter periods, respectively. The period of high correlation (Fig. 7b) exhibits a strongly zonal dipole in SLP. On the contrary, the period of low correlation was characterized by a "low-zonal" NAO dipole (Fig. 7a) and an increase in the southerly component of surface winds (Fig. 7a, wind rose). Under this situation, a positive NAO phase does not necessarily imply a comparable increase in the frequency of the westerlies due to the enhanced contribution of the meridional wind component on the wind direction. This result suggests that a decoupling between the WI and the NAO indices of winter circulation is a potential indicator of departures of the NAO dipole from zonality (i.e. migrations of the centres of action), and consequent changes in the NAO-related climatic signatures. In the light of this weak link between the WI and the winter indices of Paris-London and the NAO of Luterbacher et al. (2001a) for the 1700s–1730s and 1760s–1790s, one could argue that these might also be periods of low zonality of the NAO pattern, a possibility previously suggested by indirect measures (Jacobeit et al. 2003).

In summer, the corresponding SLP distribution for the period of strong WI-NAO linkage (1945–1974, Fig. 7d) displays the canonical pattern of the summer NAO (Folland et al. 2009), while the period of low correlation (1915–1944) was characterized by a strong dominance of the Greenland centre of variability and reduced variability over the UK (Fig. 7c). Note that the northern centre of action of the summer NAO (placed over southern Greenland) is not captured by the summer WI (cf. Fig. 3, left panels and Fig. 7d). Instead, the WI is strongly linked to SLP fluctuations over the UK, near the southern centre of variability of the summer NAO. Using data from 1900, Bladé et al. (2012) show that the centres of action of the summer NAO suffer from non-stationarity and that the spatial expression of the summer NAO became weaker prior to 1935. Therefore, one could expect that weak



◀**Fig. 7** Seasonal mean SLP fields (hPa) regressed onto the first EOF of SLP anomalies over the Euro-Atlantic region for: **a** winters of 1871–1900; **b** winters of 1901–1930; **c** high-summers of 1915–1944; **d** high-summers of 1945–1974. The explained variance is shown in the *upper right corner*. *Solid (dashed) contour lines* indicate positive (negative) values corresponding to SLP anomalies for 1 standard deviation of the PC1 time series (the 0 contour is omitted). Panels from **a** to **d** also show the seasonal frequency of daily wind direction (in % of the total number of days) for the given 31-year period in a wind rose with binsize of 10°. *Grey shading* of each directional bin is proportional to the strength of the regional mean zonal wind over the English Channel (with light-to-dark degrading denoting low-to-high values). See text for details

summer correlations between WI and the summer NAO highlight decoupling periods between the UK and Greenland centres of action of the summer North Atlantic circulation. A similar situation of low WI-NAO summer correlation was repeated during 1850–1880 (Fig. 6d), a period with previous evidence of low vorticity and small meridional pressure gradients over the English Channel (Jacobeit et al. 2003). This is consistent with the hypothesis that fluctuations of the summer NAO would be decoupled from the summer WI during periods with weak activity of the UK centre. Unfortunately, neither the reconstructed summer NAO of Folland et al. (2009) nor the summer Paris-London index correlate significantly with the summer WI before the instrumental period, which prevents the identification of potential decoupling periods between the WI and the summer NAO further back in time.

Our results suggest that in both seasons, the WI seems to be a useful indicator of potential non-stationarities of the NAO dipole. We note however that there are no clear changes in the frequency of the westerlies (WI) or in the strength of the zonal wind (NAO) linked to the periods of high/low correlation (e.g., cf. Fig. 7c, d). This means that a period of high (low) WI does not necessarily involves a strong (weak) correlation with the NAO. This also seems to be true for the circulation-climate relationship, as previously described for the linkage between the strength of the circulation and the magnitude of the correlation between circulation and temperature in Europe (Pozo-Vázquez et al. 2001; Slonosky and Yiou 2002).

6 Conclusions

The WI provides the longest North Atlantic circulation index currently available and assembled exclusively from direct weather observations. The analysis of the WI record indicates that the frequency of the westerlies in the English Channel has not suffered large long-term changes for the past three centuries, and that recent decades are not unprecedented in the long-term context. The climatic signal of the WI in precipitation reveals increased precipitation in north and central Europe and drier conditions in the Mediterranean during periods of enhanced frequency of westerlies. The WI shows a significant year-round signature on precipitation in large areas of Europe that is missed by the NAO indices. The fact that WI has a significant precipitation signal through the year is relevant because a number of climate proxies are especially sensitive to the precipitation anomalies in specific seasons. The WI also captures temperature anomalies over Greenland, the British Isles and, depending on the season, different European regions, which sometimes differ from the regions affected by the NAO. A year-round temperature signature of the WI is limited to the British Isles and a small part of central Europe, and this seems to be related to the absence of wind force information in the WI. Therefore, because of its directional nature, the WI provides additional and complementary climatic information to that of the station-based NAO.

Comparison of the WI with instrumental indices such as several NAO records and the Paris-London index resulted in overall good agreement, except for the summer Paris-London index before the instrumental period and station-based summer NAO indices biased to the winter definition of the NAO. There was also a generally poor correlation of the WI with proxy-generated indices of the NAO, with the exception of the winter NAO reconstruction of Luterbacher et al. (2001a), which includes instrumental data in its record. Therefore, the WI is associated with both the winter NAO and the high-summer NAO and hence, it can be considered a climatic indicator of past atmospheric fluctuations over the northeastern North Atlantic-European region. However, the comparison of the WI with the NAO also reveals decoupling periods between the frequency and the intensity of the zonal flow at several time scales and identifies possible causes of non-stationarity in the relation between the atmospheric circulation over the North Atlantic and the local climate variability. The assessment of the most recent of these events strongly suggests that decoupling periods are related to spatial shifts in the configuration of the NAO dipole (winter) and changes in the disposition of its centres of variability (summer). This would temporarily modify the “canonical” patterns of temperature and moisture advection over Europe associated with the North Atlantic pressure dipole, making the NAO indices based on paired pressure differences or on proxy information limited indicators of the North Atlantic atmospheric circulation during these periods. The periods of decoupling of WI with other independent instrumental climatic indices might further help to define more clearly these periods and the associated anomalies. Equally, unless this instability is considered, reconstructions based on the local temperature or precipitation might not faithfully capture the details of past North Atlantic atmospheric circulations. The causes and consequences resulting from the periods of disconnection in the pre-instrumental record between the WI and NAO are more difficult to assess. Currently, further analyses, including numerical modeling of the WI in paleoclimate simulations, are underway to deepen the understanding of the long-term variability of the WI impacts and of the relation between the WI and the NAO.

In summary, its high-temporal resolution, instrumental nature, ability to capture all-year climatic signal and the fact that it is a direct wind measure make the WI an excellent benchmark for proxy calibration, since it reduces uncertainties associated with other circulation indices derived from proxy evidence. Finally, it is important to note that there are many other areas of the globe where it would be possible to assemble WI-like indices based on historical marine data. The promising results of this study will hopefully stimulate future developments in this direction.

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