# A review of non-stationarities in climate variability of the last century with focus on the North Atlantic–European sector

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# A review of non-stationarities in climate variability of the last century with focus on the North Atlantic–European sector

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## 1. Introduction

To understand processes in the climate system and to project future climate change, analysis and characterization of the relationships between different components of the system through time are necessary. In many instances it is assumed that the statistical properties of data series remain the same through time. Examples include statistical downscaling of future climate change, hydrological statistics and modelling, and ecosystems research and planning. It implies that the variable under consideration (e.g. precipitation) has a time-invariant probability density function (pdf), whose properties can be estimated from the observational record. It corresponds to the idea of stationarity, i.e. that natural systems fluctuate within an unchanging range of variability. In climate research stationarity usually refers to a process which is weakly stationary, in which case the mean and autocovariance function of the data series are constant through time (von Storch and Zwiers, 1999; Wilks, 2006). Thus, different time slices of a stationary data series can be regarded as having the same underlying mean, variance, and covariances. Furthermore, the correlations between stationary series of different variables only depend on their relative positions in the series, and not their absolute positions in time (Wilks, 2006).

However, there is a general scientific agreement that climate is fundamentally non-stationary (Milly et al., 2008; Lins and Cohn, 2011). Non-stationarities occur on all spatial and temporal scales. On the intra-annual time scale, the annual cycle of temperature in the extra-tropics is an example for non-stationarity. Inter-annual variations in the annual cycle, i.e. variability of seasonality, can be seen in many climate time series (Pezzulli et al., 2005). Time-variant climate characteristics on the intra-annual to inter-decadal time scales are often induced by changes of the effects of the atmospheric circulation on the climate of a specific region. The changes of the atmospheric circulation are measurable for instance as changes of the frequencies and phases of atmospheric patterns. In addition, the strength and the spatial location of the pattern-related anomaly centres can change, as well as the within-pattern characteristics, like the thermal and thermo-dynamic properties of the patterns. Non-stationarities on intra-annual to multi-decadal time scales can also be observed in the preferred large-scale modes of climate variability, such as the El Niño–Southern Oscillation (ENSO), the North Atlantic Oscillation (NAO), and the Pacific–North American (PNA) pattern. These modes of variability result from the dynamical non-linear characteristics of the atmospheric circulation and through interactions with the land and ocean surfaces. On the centennial and longer time scales, non-stationarities are caused by natural internal variability as well as by external factors, such as orbital, solar irradiance, volcanic or anthropogenic forcing. Of what kind and how strong non-stationarities are or will be due to anthropogenic climate change still remains a great challenge to understand.

Many considerations on climate variability and climate change rely on analyses of the large-scale atmospheric circulation and its link to regional patterns of temperature, precipitation and other climatic variables. Thus, in the present paper we concentrate on non-stationarities representing major modifications of the relationships between atmospheric circulation and climate. Non-stationarities emerge from substantial modifications of the atmospheric circulation, which lead to significant changes of regional climate characteristics, like regional temperature and precipitation patterns. We centre our analysis on the Northern Hemisphere, particularly on the North Atlantic–European sector. Since we focus on the atmosphere–climate link, we discuss phenomena associated with the long-term heat and energy memory of the oceans only briefly. Furthermore, major modes associated with oceanic variability like the Pacific Decadal Oscillation (PDO, Mantua et al., 1997) and the Atlantic Multi-decadal Oscillation (AMO, Schlesinger and Ramankutty, 1994) are not separately discussed. Rather they are taken into consideration within the discussion of physical mechanisms underlying non-stationarities of atmospheric circulation and climate. We focus mainly on the 20th century, because of the availability of observational data.

First, in Section 2, we look at non-stationarities associated with the major modes of climate variability. We reviewed the literature regarding the occurrence and magnitude of non-stationarities in these variability modes, as well as the physical mechanisms causing the non-stationarities. We analysed the associated changes in the surface climate like modifications of the temperature and precipitation patterns. As major modes of climate variability we selected the El Niño–Southern Oscillation (ENSO) phenomenon, the dominant mode of inter-annual climate variability observed globally. Non-stationarities of Northern Hemisphere atmospheric circulation and climate are addressed by means of the Pacific–North American (PNA) pattern and the North Atlantic Oscillation (NAO). Variability of the East Asian climate is not considered in the present study. It should be noted that modes and patterns and their corresponding indices only provide a simplified description of climate variability. Consequently, the index time series just give the information on the temporal occurrence of the associated spatial patterns. The analysis of the large-scale modes of climate variability is followed by a detailed analysis of non-stationarities with an explicitly regional focus in Section 3. We took the North Atlantic–European area as an example to investigate non-stationarities in the dominant atmospheric patterns and their impact on regional climate over the European area. Beside large-scale modes of atmospheric variability, atmospheric circulation types (CTs), derived by various circulation type classifications (CTCs) are used to investigate atmospheric circulation dynamics and their effects on surface climate in the North Atlantic–European area. CTCs arrange the continuum of individual states of the atmospheric circulation into disjunct classes, utilising varying similarity measures and classification approaches to derive classifications that combine maximum homogeneity within and maximum heterogeneity among CTs (Huth et al., 2008; Philipp et al., 2010). We address the consequences of changes in the frequencies of the CTs as well as within-pattern modifications. Then, in Section 4 we discuss the main findings and highlight the consequences of circulation-climate non-stationarities for research studies which focus on climate variability and climate change or which make use of such information. We draw some conclusions and give a perspective on possible future developments.

## 2. Non-stationarities in large-scale spatial patterns of climate variability

### 2.1. ENSO

#### 2.1.1. The ENSO phenomenon

One of the most prominent sources of inter-annual variations in weather and climate around the world is the ENSO phenomenon. ENSO is an inter-annual coupled ocean–atmosphere oscillation occurring in the tropical Pacific. The sea surface temperature (SST) pattern associated with ENSO and its time series as derived from Principal Component Analysis (PCA) can be seen in Fig. 1. The atmospheric part, expressed by the Southern Oscillation (SO), is principally a seesaw (or standing wave) in atmospheric mass involving coherent exchanges of air between the Eastern and Western Hemispheres centred in tropical and subtropical latitudes, with centres of action located over Indonesia and the tropical South Pacific Ocean (Trenberth and Caron, 2000). On the intra-seasonal time scale, the power spectrum for the atmospheric ENSO principal component (PC) shows a significant peak near 20 days (Feldstein, 2000). On the inter-annual time scale, the cycle of the SO has ranged from 2 years to 6 years or longer (Trenberth and Shea, 1987). Associated with the large-scale variations in the equatorial trade wind systems are fluctuations in SST. ENSO can be regarded as an irregular oscillation between two opposite phases: El Niño and La Niña. The Walker circulation weakens during El Niño and strengthens during La Niña years (Philander, 1990). An ENSO event occurs roughly every 3–5 years, but both the recurrence interval and the strength of events exhibit strong decadal variability (Neelin et al. 1998; Diaz et al., 2001).

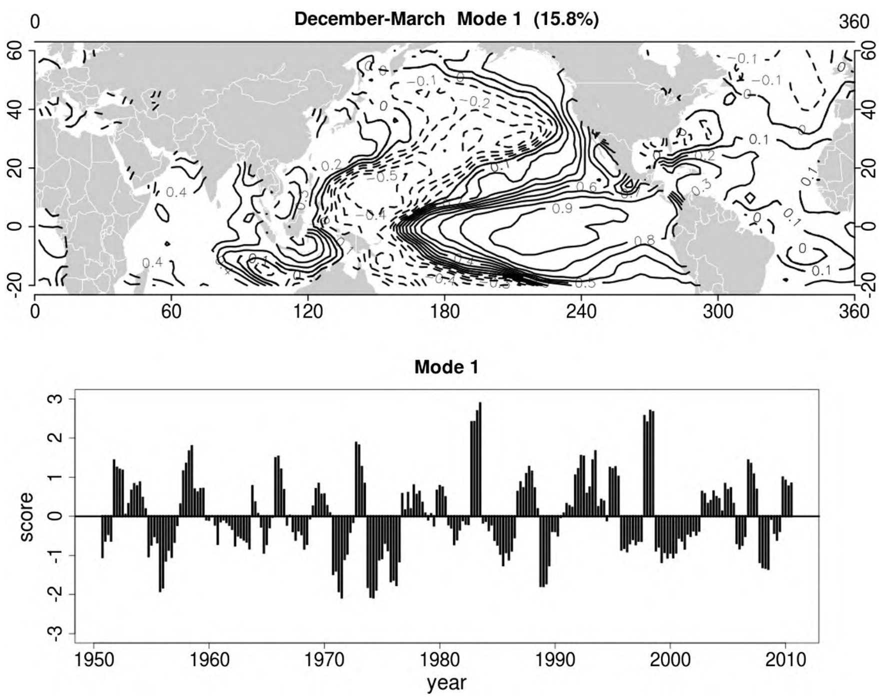


**Fig. 1.** El Niño–Southern Oscillation, represented by the first mode from VARIMAX-rotated PCA of winter (December–March) area-weighted sea surface temperatures between 20°S and 60°N. The time series shows the corresponding monthly PC scores from December 1950 to March 2010. PCA is based on standardised monthly anomaly fields from the ERSST dataset (Smith et al., 2008). Ten PCs were kept with 66.2% of total variance explained. Number in parentheses in title: percent of variance explained by the first mode.

Correlations of the Southern Oscillation index (SOI) with annual mean pressure shows statistical significance widespread over the globe, with maximum values over the Pacific–North American sector as well as over the southern oceans from Australia to the South Atlantic (Trenberth and Caron, 2000). Concerning precipitation, Ropelewski and Halpert (1987) identified 17 global core regions that appear to have a clear ENSO–precipitation relationship. Global variations in the tropical rainfall patterns are related to ENSO, most pronounced in the western and central equatorial Pacific, including most of the Australian subtropics, northern South America, eastern equatorial Africa, southern India and Sri Lanka. Mid-latitude precipitation-ENSO links were found for all major land masses of the Southern Hemisphere, for the Northern Hemisphere over the Indian subcontinent and parts of North America (Ropelewski and Halpert, 1987). An impact on temperature and precipitation was also observed over Europe (Brönnimann et al., 2007) and the Mediterranean area (Mariotti et al., 2002). With respect to temperature, Halpert and Ropelewski (1992) found highest SO influences in the tropics, most regions with relationships during both phases of the SO. Outside of the tropics, the authors identified relationships between the SO and temperature over northwest North America during both phases, over Japan during cold events, and over the western Mediterranean area during warm events. ENSO influences on climate are highly seasonal. Besides strong influences during the peak phase of El Niño in boreal winter, lagged correlations exist. After decaying in spring, correlations with surface climate in the summer following El Niño are most robust over the Indo-Northwestern Pacific region due to ENSO-induced changes of air–sea interactions in this region (Lau and Nath, 2003; Kosaka et al., 2013).

*2.1.2. Dynamics of ENSO variability*

Important characteristics of the equatorial Pacific climate system are given by its seasonal cycle and its annual mean state (Dijkstra, 2006). The seasonal cycle shows maximal (minimal) east–west equatorial SST contrast and southeast trades in the eastern part of the basin in October (April). The annual mean state is characterised by the zonal contrast between the western Pacific warm pool and the cold tongue in the eastern Pacific (Mitchell and Wallace, 1992; Dijkstra, 2006). The most dominant feature of the SO is the standing wave centred south of the equator that produces the seesaw in sea-level pressure. The seasonal cycle and the annual mean SST pattern seem to be related to El Niño variability. Thus, there is a specific sequence of the El Niño events termed by Rasmusson and Carpenter (1982) the peak, mature and transition phases of the El Niño events. This sequence has the tendency to phase lock to the seasonal cycle which manifests in the tendency of ENSO events to peak during boreal winter. By January the central Pacific SST anomalies are past their peak but still large (mature stage). A description of observed ENSO episodes can be found for instance in Rasmussen and Wallace (1983) and specifically for the 1990 to 1995 event in Trenberth and Hoar (1996). Aspects of the inter-annual evolution of the SO were addressed for example by Trenberth and Shea (1987) and Wallace et al. (1998).

The delayed oscillator theory of ENSO (e.g. Battisti and Hirst, 1989) can be used to explain the temporal evolution of an El Niño warm event and the following cold event. In general, ENSO theory involves slow sub-surface ocean dynamics, the thermodynamics of SST, and atmospheric feedbacks (Neelin et al. 1998). Dijkstra (2006) concluded that the ENSO phenomenon can be understood within a weakly nonlinear framework of equatorial ocean–atmosphere interaction. The oscillatory nature of ENSO derives from an internal mode of variability of the coupled system involving the coupled feedbacks and ocean wave dynamics. Mechanisms to explain the irregularity in the occurrence of ENSO events include deterministic chaotic behaviour, non-normal growth, and impact of small-scale processes usually referred to as noise (Dijkstra, 2006). For a detailed discussion on the physics of El Niño variability, the reader is referred to Neelin et al. (1998) and Dijkstra (2006).

When considering the spectrum of ENSO, energy is also found at lower frequencies, in particular in the decadal to inter-decadal range. The spatial pattern based on low-pass-filtered data is similar to the inter-annual ENSO pattern, except that it is broader in scale in the eastern equatorial Pacific and has enhanced magnitude in the North Pacific relative to the tropics. Thus, the pattern is also similar to the Pacific Decadal Oscillation (PDO, Liu and Alexander, 2007). With respect to the origin of the variability on decadal and longer time scales, two conceptual frameworks exist: in the classical ENSO idea, inter-decadal variability is regarded as a self-exiting oscillation generated by a positive ocean–atmosphere feedback associated with extra-tropical Rossby waves. The other theory comprises the delayed oscillator paradigm with stochastic forcing (Liu, 2012). Regardless of the proposed theory, mechanisms can be distinguished into those that attribute a dominant role to the tropics itself, and those that involve mid-latitude dynamics (Dijkstra, 2006; Liu, 2012). Diaz et al. (2001) attributed the connection between low frequency changes in tropical SST, ENSO and decadal scale changes in the general atmospheric circulation to a complex interplay between the canonical ENSO system, slow changes in SST in the Indo-Pacific area over the last century, and long-term changes in the atmospheric circulation itself. However, with respect to the time scale of the Pacific decadal and multi-decadal variability, it is still unclear what specific mechanisms are responsible (Liu, 2012). An overview on ENSO variability and teleconnections of the last decades and in the future is given by Christensen et al. (2013). An overview of recently established central equatorial El Niño specifications can be found for instance in Larkin and Harrison (2005, "Dateline" El Niño) and Ashok et al. (2007, "El Niño Modoki").

#### 2.1.3. Non-stationarities

For the 20th century low-frequency changes in the behaviour of ENSO in the context of changes in global SST patterns as well as abrupt and sustained anomalies of the coupled ocean–atmosphere system since the mid-1970s have been noted by Diaz et al. (2001). The authors found that both the recurrence interval and the strength of ENSO events exhibit considerable decadal variability. An analysis of ENSO variability over the past seven centuries based on tree-ring chronologies by Li et al. (2013) indicates that the highest ENSO variability occurred in recent decades. Also Messié and Chavez (2011) identify shifts in the tropical and North Pacific low-frequency SST variability, located around 1925, 1947, 1976, and 1998. They note that the fluctuations are consistent with PDO transitions to a warm regime in 1925, to a cold regime in 1947, and subsequently to a warm one in 1976, again. The SOI showed high values from 1900 to 1920 (Trenberth and Shea, 1987) and diminished from about the 1920s to 1950 (Trenberth and Caron, 2000). Subsequently, the SOI, Darwin mean sea level pressure, the NINO3 index, and the equatorial zonal wind-stress all reached record values in 1977–2006 (Power and Smith, 2007). The strongest warm events in the record occurred recently, during 1982–1983 and 1997–1998 (Diaz et al., 2001). Burgers and Stephenson (1999) argued that the predominance of El Niño events compared to La Niña events during 1950–1997 is related to non-linearity of ENSO, evidenced in high skewness of the SSTs in the upwelling region in the eastern equatorial Pacific. Wittenberg (2009) also points to nonlinear effects as a source for decadal-scale Pacific SST variations. Power and Smith (2007) noted that the values since 1977 show a distinct non-stationarity, i.e. that the values are unlikely to have come from the same pdf as the earlier values. The authors discussed this change in the context of global warming, namely that global warming weakens the Walker Circulation and warms the tropical Pacific Ocean. However, if the statistics are adjusted to the new mean, it is considered that global warming has little impact on tropical ENSO-driven variability about the new mean state (Power and Smith, 2007). Qian et al. (2011) analysed the prolonged 1990–1995 El Niño event and found that it was partly caused by ENSO inter-annual variability, but also by a change in the background mean state. The amplitude of ENSO inter-annual variability was found to have increased by about 30% since the late 1930s. The annual cycle shows large year-to-year variations, and the amplitude decreased by 14% between the late 1940s and 2007 (Qian et al., 2011). Wang (1995) determined that the atmospheric circulation characteristics of the El Niño onset differ substantially before and after the late 1970s due to inter-decadal changes of the background state.

With respect to relationships of ENSO with precipitation and temperature, various non-linearities have been recorded. For instance, large shifts in the conditional precipitation distribution as a function of the SO phase as well as considerable spatial variations in the typical SO-related precipitation patterns in some regions were found by Ropelewski and Halpert (1996). Halpert and Ropelewski (1992) found for southeast Africa and southeast Asia–India that the temperature response during cold event years appears to begin earlier than the response during warm event years.

In order to investigate non-stationarity in ENSO teleconnections, Van Oldenborgh and Burgers (2005) looked at decadal variations in the strength of ENSO teleconnections to precipitation using station data over the whole globe. The authors found a coherent signal only in April–June in northern Europe and in July–September in southern Australia and the mid-western United States. In contrast, Diaz et al. (2001) noted that the ENSO teleconnections have been non-stationary throughout the 20th century in various regions including the southeast United States, the South Asian/Indian monsoon region, the Argentinean sector, and the Caribbean region. Thus, for instance, the south-eastern parts of the United States showed a strong correlation between wintertime precipitation and ENSO in the last half of the 20th century, but near-zero correlation in the period before. Further aspects on non-stationarities in the ENSO–North American climate teleconnections are discussed in context with the PNA pattern in Section 2.2.

For the Amazon region of South America, Espinoza Villar et al. (2009) noted a change in July–August/September–November total rainfall at the end of the 1970s/beginning of the 1980s. The authors assumed that the decreased rainfall totals after that date can be attributed to the warming of the tropical Pacific. During this period of more El Niño dominated conditions with a generally lower SOI, increased subsidence and a northward-shifted inter-tropical convergence zone over northeast Brazil and the Amazon Basin and a southeast shift in the southern Atlantic convergence zone occurred (Haylock et al., 2005).

With respect to the ENSO–rainfall relationship over Australia, Suppiah (2004) analysed the period 1889–1996 and found that a significant positive relationship existed for most of the time. However, he also found two periods of diminished relationship between the SOI and Australian rainfall: the 1920s and 1930s, and the period after the mid-1970s. The former period was characterised by a strong decrease in rainfall and a small increase in the SOI, whereas the recent period was characterised by a strong decrease in the SOI and a small increase in rainfall. Thus, increases in rainfall during the 1980s and the 1990s and decreases in the SOI have weakened their relationship resulting in more rainfall for a given SOI after 1973 (Suppiah, 2004)

The near-constant negative correlation between the Indian summer monsoon rainfall and ENSO for most of the 20th century has recently shown a significant weakening (Diaz et al., 2001). However, Wang et al. (2008) found that at the same time the overall coupling between the Australian–Asian Monsoon system and ENSO has been enhanced since the late 1970s. Enhanced positive correlations between ENSO and the western North Pacific, East Asian, and Indonesian monsoons during the developing, maturity, and decaying phases of ENSO emerged, regardless of the weakening of the Indian monsoon–ENSO correlation. Wang et al. (2008) argued that these inter-decadal changes are attributed to increased magnitude and periodicity of ENSO and the strengthened monsoon–ocean interaction. The enhanced ENSO variability has increased the strength of the monsoon–warm–pool interaction and the Indian Ocean dipole (Saji et al., 1999) SST anomalies. This resulted in a strengthening of the summer westerly monsoon across South Asia and thus in a weakening of the negative linkage between the Indian summer monsoon rainfall and the eastern Pacific SST anomaly. On the other hand, the amplified ENSO forcing and the induced monsoon–ocean interaction have reinforced the impact of ENSO on the East Asian, western North Pacific, and Indonesian monsoons (Wang et al., 2008). Xie et al. (2010) and Chowdary et al. (2012) point to the importance of tropical Indian Ocean SST anomalies for the ENSO–monsoon teleconnections. After the mid-1970s a slow decay rate during summer following an El Niño event induces a more robust response over the tropical Indian Ocean. The strong Indian Ocean SST response leads to a pronounced development of anticyclonic circulation anomalies over the Northwest Pacific and East Asia.

Non-stationary ENSO teleconnections were also found for the Euro-Atlantic winter climate. Greatbach et al. (2004) showed that in the post 1970s period the ENSO signal in the form of a PNA-like wave train has intensified and is located further poleward due to changes in the spatial pattern of the tropical forcing. Mariotti et al. (2002) determined substantial modifications of the ENSO teleconnections to Euro-Atlantic precipitation during the 20th century. Thus, significant positive values with western Mediterranean rainfall became visible for the autumn season starting only from the early 1940s. For spring, significant positive values were found early in the century and after the late 1960s. A lack of significant correlation for either season was observed during the period 1925–1940, coinciding with the period of diminished SO. In contrast, on multi-decadal time scales, no significantly non-stationary behaviour of the influence of ENSO on European climate was found by Brönnimann et al. (2007) in a 51-year moving window analysis of the last five centuries.

Non-stationarities are an intrinsic part of the ENSO system. They are mainly associated with the dynamics of ENSO variability. Strong modulations of the ENSO behaviour can arise solely from stochastic processes associated with ENSO's interannual time scale and seasonal phase-locking as shown by Wittenberg (2009) using a 2000 years control simulation of a global coupled General Circulation Model. Non-stationarities of ENSO can be related to modulations of the coupled atmosphere-ocean system as highlighted for instance by Chowdary et al. (2012) for the ENSO–Indo-western Pacific teleconnections. Furthermore, separate decadal climate modes might impact on the ENSO system, causing non-stationarities (e.g. associated with the AMO, Kang et al., 2014). However, ENSO characteristics can also be altered by external forcing like large tropical volcanic eruptions (Li et al., 2013) and anthropogenic climate change (Power and Smith, 2007). Because the influence of ENSO on regional climate shows significance widespread over the globe, non-stationarities associated with ENSO can be found in many regions. In general, factors which govern the strength of the ENSO teleconnections include changes in the evolution of the SST pattern during ENSO, changes in the ENSO's temporal behaviour, changes in the mean base climatology and its modulation of the teleconnection processes (Diaz et al., 2001).

### *2.2. Pacific–North American area*

#### *2.2.1. The Pacific–North American (PNA) pattern*

The PNA pattern is a major mode of atmospheric variability over the North Pacific and North America especially in winter. The pattern includes a north–south seesaw in the central Pacific together with centres of action over western Canada and the south-eastern United States (Wallace and Gutzler, 1981). It can be seen as a Rossby wave train with these four centres of action. The pattern and its time series, obtained by applying rotated PCA to winter (December–February) geopotential heights of the 500 hPa level, are shown in the lower panel (Mode 2) of Fig. 2. The intrinsic time scale of the PNA pattern is about 2 weeks (Franzke et al., 2011), but it exhibits substantial variability on inter-annual and inter-decadal time scales (Wallace and Gutzler, 1981).

The PNA pattern is highly correlated with regional temperature and precipitation over the United States (Leathers et al., 1991). In winter, there are areas of high negative correlation coefficients with surface temperature over the south-eastern United States and high positive values over the northwest. In the centre of the country, no significant correlations occur between the PNA index and temperature. During a positive phase of the PNA pattern, there is enhanced ridging over the western North America and a southerly displaced and anomalously deep trough over the southeast. Anomalous precipitation patterns associated with the positive phase of the PNA pattern include below average precipitation over the upper Mississippi Valley and the northern Rocky Mountains. With negative PNA index values, the polar front jet is pushed northwards of its mean position, allowing the more frequent intrusion of air masses from the south and from the west into the above mentioned regions. This results in increased precipitation in these areas (Leathers et al., 1991).

#### *2.2.2. Dynamics of PNA variability*

Mechanisms which control the emergence and maintenance of the PNA pattern are linear dispersion of a Rossby wave emanating from anomalous tropical heating, barotropic amplification due to the zonal asymmetry of the climatological flow, and driving by synoptic scale transient eddy vorticity fluxes (Franzke et al., 2011). Thus, the positive (negative) mode of the PNA pattern is triggered by intensification (reduction) of convection over the tropical west Pacific and weakening (enhancement) of convection over the Indian Ocean (Deser et al., 2004; Johnson and Feldstein, 2010; Franzke et al., 2011). On intra-seasonal time scales, the PNA pattern is strongly influenced by the Madden–Julian Oscillation and fluctuations in the east Asian jet stream (Higgins and Mo, 1997; Johnson and Feldstein, 2010). On inter-annual time scales, ENSO impacts on the PNA pattern through modifications of the Aleutian low. Observations (e.g. Trenberth, 1990) and modelling studies (e.g. Blackmon et al., 1983; Graham, 1994) found a link between tropical SSTs, especially in terms of a negative SO-index (El Niño conditions), and a deeper Aleutian low. The deepened Aleutian low during ENSO events results in a SST anomaly pattern with cooler SSTs in the central Pacific and warming off the coast of western North America. The SST pattern is enhanced through positive feedback effects from the extra-tropical SST anomaly itself and from changes in momentum and vorticity fluxes associated with changes in high-frequency storm tracks.

On decadal to inter-decadal time scales, coupled ocean–atmosphere interactions are discussed as the source of North Pacific climate variability. In some theories, the interactions involve the extra-tropical and tropical Pacific, connected by an atmospheric bridge and by oceanic pathways. The main ocean–atmosphere relationship of the extra-tropics is one where the changes in the atmospheric circulation induce corresponding SST anomalies. However, there are strong indications that extra-tropical SST anomalies also influence the atmospheric circulation. The upper ocean heat storage is regarded to add further persistence to the extra-tropical system (Miller et al., 1994; Trenberth and Hurrell, 1994; Deser et al., 2004). For a further discussion on the origin of Pacific inter- to multi-decadal climate variability, the reader is referred to Liu (2012) and references therein.

#### *2.2.3. Non-stationarities*

The concept of non-stationarities in the Pacific region is closely related to so-called "regime shifts" (e.g. Overland et al., 2008), a characteristic observed in the low-frequency climate variability. Analysis of the intensity of the Aleutian low reveals significantly different conditions in the winters before and after the winter 1976–1977 (Trenberth, 1990; Miller et al., 1994). According to Trenberth (1990), the centre of the Aleutian low in the period 1977–1988 was located farther east and on average over 4 hPa deeper than in the period 1946–1976. Associated with this anomaly was a warming of >1.5 °C in Alaska and a cooling of >0.75 °C in the central and western North Pacific in the later period (Trenberth, 1990). Precipitation and streamflow decreased in the western US (Miller et al., 1994). North Pacific marine ecosystems were also substantially affected (Hare and Mantua, 2000).

In addition to the major phase transition of the Pacific Decadal Oscillation (PDO) observed in 1976/1977, further regime shifts in the North Pacific have been detected. Thus, Minobe (1997) identifies multiple regime shifts around 1890, 1920s, 1940s and 1970s, associated with 50–70 year climate variability over the North Pacific and North America. He suggests that it is an internal oscillation in the coupled atmosphere–ocean system, possibly being modulated in the 20th century by external solar radiation heating. Rodionov (2004) used a sequential algorithm on the January PDO index and found abrupt regime shifts located at the

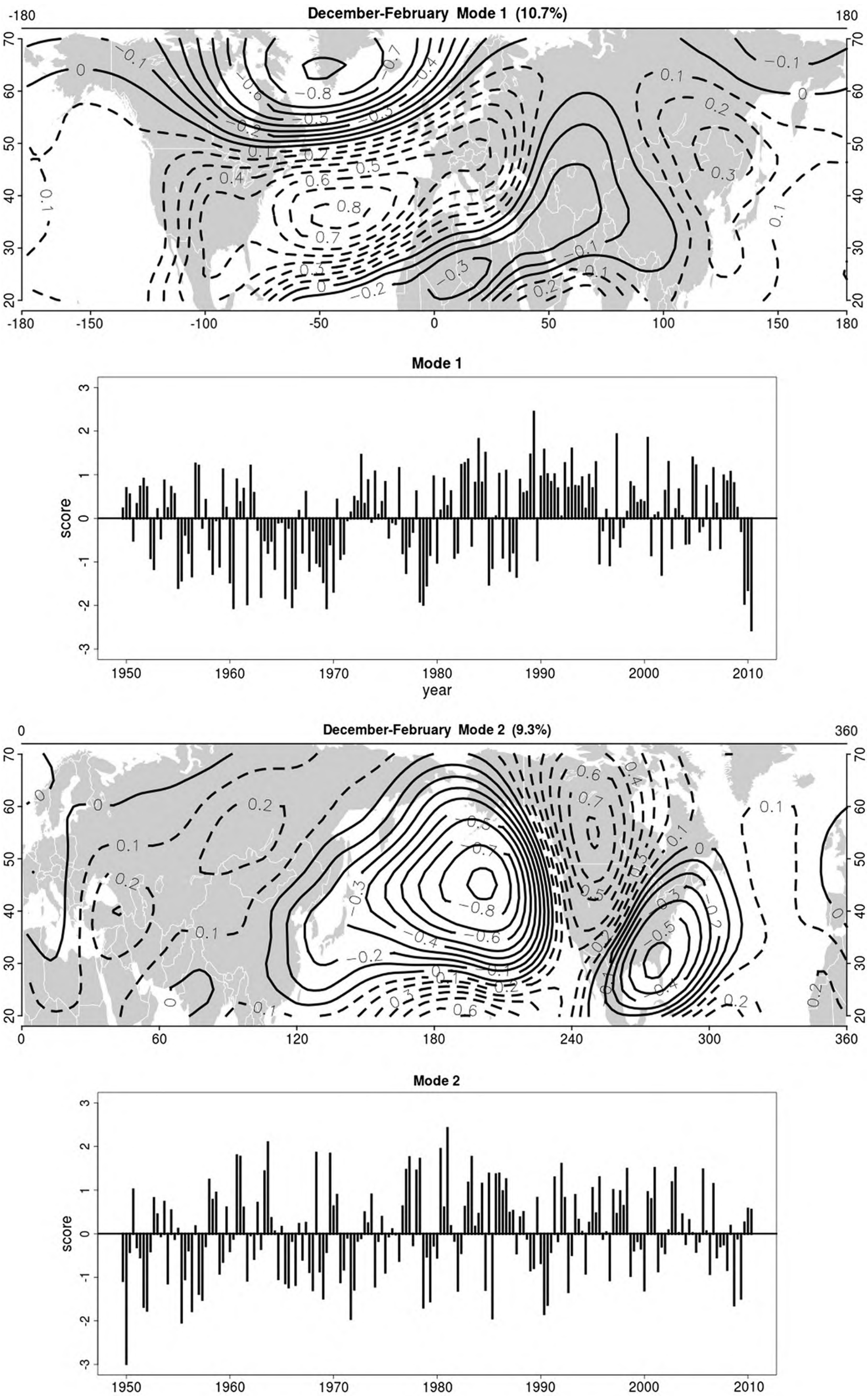


**Fig. 2.** North Atlantic Oscillation (upper figure) and Pacific–North American pattern (lower figure), represented by the first and second mode, respectively, from VARIMAX-rotated PCA of winter (December–February) area-weighted 500 hPa geopotential heights of the Northern Hemisphere between 20°N and 70°N. Time series show the corresponding monthly PC scores from December 1949 to February 2010. PCA is based on standardised monthly anomaly fields from the NCEP/NCAR reanalysis (Kalnay et al., 1996; Kistler et al., 2001). Ten PCs were kept with 73.2% of total variance explained. Numbers in parentheses in title: percent of variance explained by each mode. Note the different centring of the plots.

years 1910, 1922, 1943, 1958, 1977, and 1989. Overland et al. (2008) determined shifts in multiple climate indices of the North Pacific area around 1989 and 1998. The authors integrated these climate shifts into the picture of a long-memory process with considerable autocorrelation, which can show persistent major deviations from the long-term century-scale mean. However, Rodionov (2006) noted that the major shift in 1976 is unlikely just a manifestation of a red noise process. Johnson and Feldstein (2010) highlighted that the PNA does not consist of a single spatial pattern, but can be regarded as many spatially varying PNA-like patterns. They described the 1976/1977 regime shift as a transition from the dominance of various patterns with anomalous high pressure over the North Pacific to an epoch dominated by patterns with anomalous low pressure. Consequently, inter-decadal variability over the North Pacific involves not only changes in the strength of the Aleutian low, but also changes in the spatial structure of the Aleutian low and surrounding anomaly centres. Analysing the 1989 and 1998 shifts in the state of the North Pacific, Bond et al. (2003) identified spatial patterns in sea level pressures (SLP) and SST to be responsible for the shifts, which do not resemble PDO states, but can be expressed as a second mode of variability in the North Pacific area. The 1989 shift is attributed to the end of the dominance of the PDO and the subsequent establishment of this second mode of variability. The shift in 1998 is due to the change from a strong negative phase of the second mode to a positive one. The authors concluded that the development in recent years contrasts with PDO type variability that dominated much of the 20th century. Messié and Chavez (2011) suggest that a new and strengthening mode of climate and ocean variability has become apparent since around the 1960s in agreement with an increased persistence and frequency of El Niño Modoki events in this period and an increased variance of the North Pacific Gyre Oscillation (NPGO) since the early 1990s. The NPGO represents the second leading mode of Northeast Pacific SST anomalies, reflecting changes in the intensity of the North Pacific gyre circulations (Di Lorenzo et al., 2008).

### 2.3. North Atlantic–European area

#### 2.3.1. The North Atlantic Oscillation (NAO)

The NAO, the most prominent and recurrent pattern of atmospheric variability over the Northern Hemisphere middle and high latitudes, affects temperature and precipitation over the greater European area (Wanner et al., 2001; Hurrell et al., 2003). The NAO is associated with fluctuations in the strength of the climatological mean jet stream over the western Atlantic (Wallace and Gutzler, 1981). The sea level pressure distribution over the North Atlantic for the positive mode of the NAO includes a well-developed Icelandic Low and Azores High, associated with stronger westerlies over the eastern North Atlantic and the European continent. The axis of the westerly jet is oriented southwest–northeast and points to southern Scandinavia. In the negative NAO mode, the Icelandic Low and the Azores High are weakly developed or rarely even reversed, resulting in reduced or ceased westerlies over the eastern North Atlantic. In this case the jet axis runs west–east and points to the northern Mediterranean area (Wanner et al., 2001). The NAO is closely related to the Arctic Oscillation (AO, Lorenz, 1951), which is the leading mode of SLP across the Northern Hemisphere north of 20° latitude (Thompson and Wallace, 1998). The NAO can be seen as the regional representation of the AO in the Atlantic sector with highest similarity during the winter months from December to February (Budikova, 2012). Ambaum et al. (2001) argue that the NAO may be a more physically relevant and robust representation than the AO due to the ubiquitous appearance of the NAO in various different statistical analyses of Northern Hemisphere variability.

Although the NAO is evident throughout the year, it is most pronounced during winter, accounting for more than one-third of the total variance of the North Atlantic sea level pressure field (Hurrell and van Loon, 1997). The NAO pattern and its time series, derived from PCA of winter (December–February) 500 hPa geopotential heights of the Northern Hemisphere between 20°N and 70°N, are shown in the upper panel (Mode 1) of Fig. 2. The intra-seasonal temporal evolution of the NAO pattern can be interpreted as being a stochastic process with a timescale of 9.5 days (Feldstein, 2000). On the inter-annual and longer time scales, the NAO index shows a weakly red power density spectrum with non-significant variance peaks at biennial periods and at 6 to 10 year periods (Hurrell and van Loon, 1997; Wunsch, 1999).

In the positive phase of the NAO, precipitation is anomalously high over Scotland and south-western Norway, in case of a strong negative NAO phase, high amounts of precipitation occur in the Mediterranean area and the Black Sea (Wanner et al., 2001). Surface temperature changes associated with the NAO occur most pronounced over the northwest Atlantic and extend from northern Europe across much of Eurasia. Changes in temperatures over northern Africa and the southeast United States are also notable (Hurrell and van Loon, 1997).

#### 2.3.2. Dynamics of NAO variability

Variability of the NAO on the inter-annual to inter-decadal range involves the variability within the North Atlantic Ocean, including the tropical Atlantic and the northern polar basin (Wanner et al., 2001). The primary ocean–atmosphere interaction on intra- to inter-annual time scales is associated with the NAO affecting the ocean through substantial changes in surface wind patterns and latent and sensible heat exchanges. The greatest influence on latent and sensible fluxes is associated with the meridional wind component, and to a lesser extent to a more zonal wind, indicating increased storminess (Cayan, 1992). The main pattern of covariability is given by the NAO and the North Atlantic SST tripole. (Wanner et al., 2001; Czaja and Frankignoul, 2002; Gastineau et al., 2013). Besides, Czaja and Frankignoul (2002) identified an association between the North Atlantic Horseshoe, a large-scale Pan-Atlantic SST pattern, and the early winter NAO when the SST pattern leads the NAO by up to 6 months. The connection is based on the instantaneous response of the NAO in terms of its negative state to persistent Atlantic SST anomalies. Remote influences of tropical oceanic variability on the NAO has also received attention in recent literature and are discussed for instance by Xie and Carton (2004) regarding the tropical Atlantic Ocean variability and by Toniazzo and Scaife (2006) with respect to the tropical Pacific.

For the interpretation of the NAO variability on decadal to multidecadal time scales, the dynamics of the thermohaline circulation and gyre circulation have to be considered. It involves complex underlying dynamical processes (see Wanner et al., 2001; Liu, 2012 for a discussion). A main question is related to the time scale of a specific decadal to inter-decadal variability. For instance, the Atlantic Multidecadal Oscillation (AMO) showed cool phases in the 1900s to 1920s and 1960s to 1980s, while a warm phase occurred in the 1930s to 1950s (Knight et al., 2006). In a recent study, Gastineau et al. (2013) suggested that the Atlantic Meridional Overturning Circulation (AMOC), which largely drives the AMO, is also a driver of the North Atlantic Horseshoe pattern, with lead times of the AMOC of 9 years. Álvarez-García et al. (2011) analysed the SST tripole pattern and found that the subpolar and tropical poles present quasi-decadal variations with a period of about 9 years. The variations on this time scale are associated with the atmospheric east Atlantic pattern (see below). The SST centre of action in the western mid-latitudes is characterised by a longer time scale of about 14 years and seems to be related to the NAO pattern.

#### 2.3.3. Non-stationarities

The NAO exhibits considerable long-range persistence as well as strong decadal variations and trends (Stephenson et al., 2000; Wanner et al., 2001). The winter NAO index shows the strongest positive anomalies from 1900 to 1930 and 1980 to 1989 and the strongest negative anomalies in 1870 to 1989 and in the 1950s to 1970s (Slonosky et al., 2000; Stephenson et al., 2000). There are different ways to obtain the spatial structure as well as the time series index of the NAO. Commonly, NAO representations are station-based (e.g. Jones et al., 1997) or are derived from PCA (see e.g. upper panel of Fig. 2), with either method having its strengths and limitations (for a discussion see e.g. Hurrell and Deser, 2009). Pozo-Vázquez et al. (2001), using wavelet transform, found that the temporal modes of NAO variability are strongly non-stationary. During the periods 1842–1868 and 1964–1994, variations were in the band of 6–10 years, whereas in the periods 1857–1875, 1939–1946 and 1956–1966, the band of 4–6 years dominated. Between 1875 and 1940, the spectrum of the NAO was found to be almost white. Mills (2004), using a structural time series model, identified three components in the NAO time series: long swings overlain by a cyclic pattern having a period of about 7.5 years, and an irregular component that dominates the variation of the index.

There was a pronounced positive trend of the NAO index from predominantly low values during the 1960s to predominantly high values during the 1990s (Hurrell and Deser, 2009). This trend came along with an eastward shift of the NAO centers of inter-annual variability (Jung et al., 2003; Beranová and Huth, 2008). As mechanisms for the observed eastward shift, a non-linear dependence on the strength of the background flow (Luo and Gong, 2006) and changes in solar activity (Gimeno et al., 2003) were discussed. In addition, Dong et al. (2011) demonstrated that the annular mode structure of the NAO changed with respect to a deep penetration into the stratosphere and significant anomalies over the North Pacific. It points to an enhanced role of the stratosphere in inter-annual NAO variability and to strengthened relationship between the NAO and the AO since about the mid-1970s. Furthermore, enhanced greenhouse gas forcing can contribute to changes in the spatial patterns of atmospheric variability (Brandefelt, 2006). The CMIP5 (fifth Coupled Model Intercomparison Project) models simulate an increase of the NAO index in all four seasons between 1860 and 2100, with the largest increase in autumn (Gillett and Fyfe, 2013). However, since the 1990s there has been a negative trend of the NAO index, with some extreme negative excursions, being most pronounced for the winter 2009/2010 (Osborn, 2011). The high NAO variance is discussed within the context of natural variability and has also been attributed to solar variability as well as to effects from ENSO and the Quasi Biennial Oscillation (Hanna et al., 2014).

Cassou et al. (2004) suggested that the spatial position of the NAO action centres is related to the dominant sign of the NAO index. This non-linearity in NAO variability can also be seen when looking at so-called “weather regimes” of the North Atlantic area obtained by cluster analysis (e.g. Michelangeli et al., 1995; Hertig and Jacobeit, 2014a). Spatial asymmetries are evident between the two NAO regimes with a remarkable difference in the position of the pressure anomalies, in particular the eastward shift of the sub-polar anomaly in the positive compared to the negative regime. However, Hertig and Jacobeit (2014b) also found intra-mode spatial variability. In case of the NAO+ regime, the subtropical high pressure centre extended further east in the period 1970–2000, whereas it is mostly confined to the Atlantic area in the periods 1950–1969 and 2001–2010. In the NAO-regime, the southern negative anomaly centre was much stronger and more confined to the North Atlantic in the period 1970–2000. Furthermore, Vicente-Serrano and López-Moreno (2008) detected no accumulation of positive or negative NAO values in the years in which the NAO shows a pattern with a dominant eastward shift. Thus, the non-stationarities in the spatial NAO pattern probably do not result from the preference of the positive or negative phase of the NAO. Vicente-Serrano and López-Moreno (2008) also noted that the position and magnitude of the NAO anomaly centres show marked inter-decadal variability. The eastward change beginning in the late 1970s was not a unique phenomenon of NAO variability. The authors suggested the possible existence of different NAO types related to the location and surface extent of main pressure centres. Thus, for instance, Mächel et al. (1998) found during the decade 1921–1930 that the Icelandic Low was located about 6° further west than on average, and the core of the Azores High about 1° southwards. The authors also found a comparatively sudden fall in pressure of the Icelandic Low at the end of the 1980s and a fast rise in pressure in the late 1960s. During the winters 1989–1995, the Azores High reached an unusually high intensity, which was accompanied by a considerable north-eastward shift of its core.

Some studies showed that the magnitude of NAO variability on interdecadal to multi-decadal time scales changes. For instance, Slonosky et al. (2000) found that the variability of the zonal atmospheric circulation over Europe was more variable, with more extreme values and sudden changes, in the late 18th century and the first half of the 19th century compared to the 20th century. In contrast, Goodkin et al. (2008) determined enhanced multi-decadal scale variability of the NAO during the late 20th century compared with the period 1800–1850. The authors suggested that NAO variability is linked to the mean temperatures of the Northern Hemisphere with warming acting to increase NAO multi-decadal variability considerably. Beyond, Zanchettin et al. (2012) showed in a modelling study that large tropical volcanic eruptions can produce decadal and longer oceanic–atmospheric anomalies involving major changes in the phasing of the AMOC, the North Atlantic subpolar gyre and the state of the winter NAO. However, it should also be noted that, beside physical changes underlying the variability of the NAO, some uncertainty exists in the observational analyses due to data availability and due to specifics of the NAO index calculation procedures (e.g. homogenization procedures of the pressure series, normalization period used, and months included; Slonosky and Yiou, 2001).

The non-stationary and non-linear variability of the NAO has direct consequences for the relationship between the NAO and associated climate anomalies. Pozo-Vázquez et al. (2001) found a strong linear relationship between the NAO and temperature of the British Isles and southern Scandinavia. In contrast, Central European temperatures are only influenced by strong positive NAO phases, but show little changes for moderate NAO events and extreme negative NAO phases. Temperatures of the Iberian Peninsula seem to be more influenced by moderately positive NAO anomalies than by an extremely positive NAO. Diao et al. (2014) show that there exists an asymmetry in location between warm and cold temperature extremes in Europe associated with the NAO. Kenyon and Hegerl (2008) find differences in the shape of the temperature distribution between negative and positive NAO years with a much wider lower tail of the distribution for negative NAO years.

In addition to these non-linear characteristics of the NAO-climate relationships, a considerable non-stationary behaviour becomes visible in temporal correlations between the NAO and surface temperature and precipitation. For several European locations, Jacobeit et al. (2001) reported a marked decline in NAO-temperature correlations at the beginning of the 20th century. Comparable sudden decreases in correlations between NAO and temperatures around the turn from the 19th to the 20th century have also been depicted by Chen and Hellström (1999) for Sweden, and by Slonosky et al. (2001) for various European stations. Reid et al. (2001) documented an increase in correlations between NAO and temperature and precipitation in winter over Europe between 1901–1950 and 1951–1995. Beranová and Huth (2008) showed that running correlations at a large number of stations in southern Europe and in the Balkans decrease from the start of their analysis in 1958 to the mid-1970s and then increase up to the end of the century. Stations located in western, central and south-eastern Europe show an intense increase in correlation from the beginning of the 1970s to the end of the century, whereas stations of Iceland and Norway have a dramatic decrease since the mid-1970s. Concerning NAO-precipitation relationships, the authors find decreasing correlations during 1958 and 1998 for stations in southern parts of Europe, and increasing values for stations in northern Europe. Stations in the British Isles, western, central, and eastern Europe have running correlations decreasing prior to 1975 and significantly increasing after that. Comparing the correlation strength between NAO and precipitation over the Iberian Peninsula in the two periods 1958–1977 and 1978–1997, Goodess and Jones (2002) revealed higher/lower correlations in the more recent period for winter/spring. In a longer period from 1902 to 2000, Vicente-Serrano and López-Moreno (2008) also identified non-stationary correlations of the NAO index with European precipitation. The authors found that the general pattern of negative (positive) correlations in the south (north) of Europe between winter precipitation and the NAO index was preserved, but spatial pattern and magnitude of the correlations varied noticeably over time. Based on reconstructed sea level pressure and precipitation data since 1500, Pauling et al. (2006) discerned distinct non-stationarities in correlations between NAO and precipitation over Southern Spain and Morocco with significantly negative correlations prevailing during the 19th and for parts of the 17th and 20th centuries and intervening periods of insignificantly negative or even positive correlations.

## 3. Non-stationarities over the North Atlantic–European area

Although the NAO is the dominant pattern of atmospheric circulation variability over the North Atlantic, it explains only a fraction of the total variance, and most winters cannot be characterised solely by the NAO pattern. Beyond that, Jacobeit et al. (2001) noted that each NAO mode is linked to a variety of different circulation patterns over Europe implying a complex relationship with regional climate. A particular impact on regional climate may crucially depend on the local or regional details of the atmospheric circulation pattern that is forcing it, and these details may not always be adequately reflected in a simple index such as the NAO (Hurrell and Deser, 2009).

### 3.1. Non-NAO modes

Concerning large-scale modes of atmospheric variability over the North Atlantic–European domain in addition to the NAO, Barnston and Livezey (1987), among others, identified through PCA the north-south dipole East Atlantic (EA) pattern and two other patterns over the Eurasian continent (Eurasian Type 1 (EU1) and Eurasian Type 2 (EU2) patterns), which are each characterised by an east–west wave train of three centres of action. Also in teleconnection analysis of winter geopotential heights (e.g. Wallace and Gutzler, 1981), the EA pattern and the EU1 pattern can be found. The EA, EU1 and EU2 patterns and their time series as derived from PCA of winter 500 hPa geopotential heights in the period 1950–2010 are displayed in Fig. 3.

The EA pattern is significantly positively correlated with temperature over western, southern, and central Europe (Beranová and Huth, 2008). Positive correlations with precipitation are found along the western coast of Europe and the Iberian Peninsula, negative ones over the central part of Western Europe and over south-eastern Europe (Wibig, 1999; Beranová and Huth, 2008). A noticeable change in the EA pattern occurred in the second half of the 20th century. The northern extension of the southern centre of action retreated south- and eastwards, while the northern centre of action extended towards Scandinavia (Beranová and Huth, 2008). This is connected to changes in the EA-relationships with temperature and precipitation over Europe with non-stationarities occurring preferably in the 1970s (Beranová and Huth, 2008). Non-stationarities in the connections of the EA pattern with circulation patterns and precipitation over the eastern Mediterranean area in the early 1990s and western Mediterranean area in the mid-1980s were found by Hertig and Jacobeit (2013). Non-stationarities in the correlations between EA and precipitation over the eastern Mediterranean have likewise been reported by Touchan et al. (2005) who found significantly positive correlations only around the mid of the 19th and 20th centuries, while mostly insignificant correlations of varying sign alternate over most of the period since 1764.

The EU1 pattern (also referred to as the Scandinavia pattern) shows significantly negative correlations with temperature at stations in the British Isles, Denmark, and the Iberian Peninsula, while significantly positive values occur over northern Scandinavia (Beranová and Huth, 2008). Statistically significant positive correlations with precipitation occur mainly in the central Mediterranean area, negative values are located over north-eastern Europe and near the southern and eastern coast of the Baltic Sea (Wibig, 1999; Beranová and Huth, 2008). The EU1 pattern exhibits large intra-seasonal, inter-annual and inter-decadal variability. Negative phases of the pattern with below-average pressure in north-eastern Europe and the extension of the Azores high in Western Europe were observed from 1964 to 1980 and again from 1986 to 1993 (Panagiotopoulos et al., 2002). The centre of action over Spain underwent changes in its shape and position, with intensification since 1986. This induced, under the negative EU1 phase, a more westerly than northerly flow into northern, central, and southern Europe (Beranová and Huth, 2008). Significant decreases in correlation in the period 1958–1998 were found for southern Scandinavia, the British Isles, eastern Europe, and the northeast of the Balkans, whereas northern Scandinavia, Iceland, France and Italy showed an increase up to 1970 and a decrease from about 1980 (Beranová and Huth, 2008). Non-stationarities in the EU1-precipitation correlations were found for stations north of 50°N and along the Bay of Biscay with a pronounced decrease in correlations prior to 1975, recovering after 1980 (Beranová and Huth, 2008).

The EU2 pattern (also referred to as the East Atlantic/West Russia pattern or the North Sea–Caspian pattern) shows significantly positive correlations with temperature over the northern half of Europe and stations in the Alps. Significantly negative correlations were found in the southeast of Europe and the eastern Mediterranean area (Kutiel et al., 2002; Beranová and Huth, 2008). The pattern correlates significantly negative with precipitation over the eastern Atlantic and positive with precipitation over the south-eastern Mediterranean regions (Wibig, 1999; Krichak and Alpert, 2005). The EU2 pattern exhibits considerable inter-seasonal and inter-annual variability. Persistent negative phases were observed during the winters and early springs of 1969/1970, 1976/1977 and 1978/1979 characterised by enhanced precipitation in Western Europe and southerly flow over most of Europe. Pronounced positive phases of the pattern are less common (Panagiotopoulos et al., 2002). However, a positive trend of the EU2 pattern during the second half of the 20th century was found by Krichak and Alpert (2005). In addition, the authors detected notable non-stationarities in the EU2-precipitation correlation patterns between periods with low and high EU2 index values associated with differences in the orientation of the air-mass transport. The positive trend as well as the non-linearity between the positive and negative phases contributed to the observed precipitation decline over the eastern Mediterranean area during the analysed period (Krichak and Alpert, 2005).

Overall, non-stationarities are evident in all large-scale atmospheric patterns. They are related to changes in the frequency of occurrence of specific patterns and to changes in the dominance of preferred phases of the patterns. The commonly non-linear character of the phases has an important effect in this context. A second source for non-stationarity lies in changes in the structure of the patterns, ranging from changes in the mean pattern state to changes of the intensity and spatial location of the centres of action. Associated changes of the pattern-specific circulation characteristics alter the circulation-climate relationships, and consequently non-stationarities occur in regional climate influenced by the specific patterns.

### 3.2. Circulation types over the European area

Beside large-scale modes of atmospheric variability, atmospheric circulation types (CTs) derived by various circulation type classifications (CTCs), are used to investigate atmospheric circulation dynamics and their effects on surface climate in the North Atlantic–European area. 1000 hPa geopotential height composites of ten North-Atlantic European CTs derived from an objective classification (Beck et al., 2007) of daily 1000 hPa geopotential heights in the period 1950–2010 are shown in Fig. 4. Their annual occurrence frequencies are plotted in Fig. 5. From several studies investigating the relationship between CTs and surface climate in the North Atlantic–European area, it turned out that the observed changes—mainly during the 20th century—in varying surface climate parameters can only partly be attributed to corresponding variations in occurrence frequencies of CTs. Concerning trends in several climate elements observed in the Czech Republic during the second half of the 20th century, Huth (2001) and Cahynova and Huth (2009, 2010) found that trends in frequencies of certain CTs only explain parts of the observed trends in winter temperatures, whereas climatic variations in summer appear largely unrelated to frequency changes of CTs. Cahynova and Huth (2010) furthermore quantified the contribution of frequency changes of CTs to the observed climatic trends

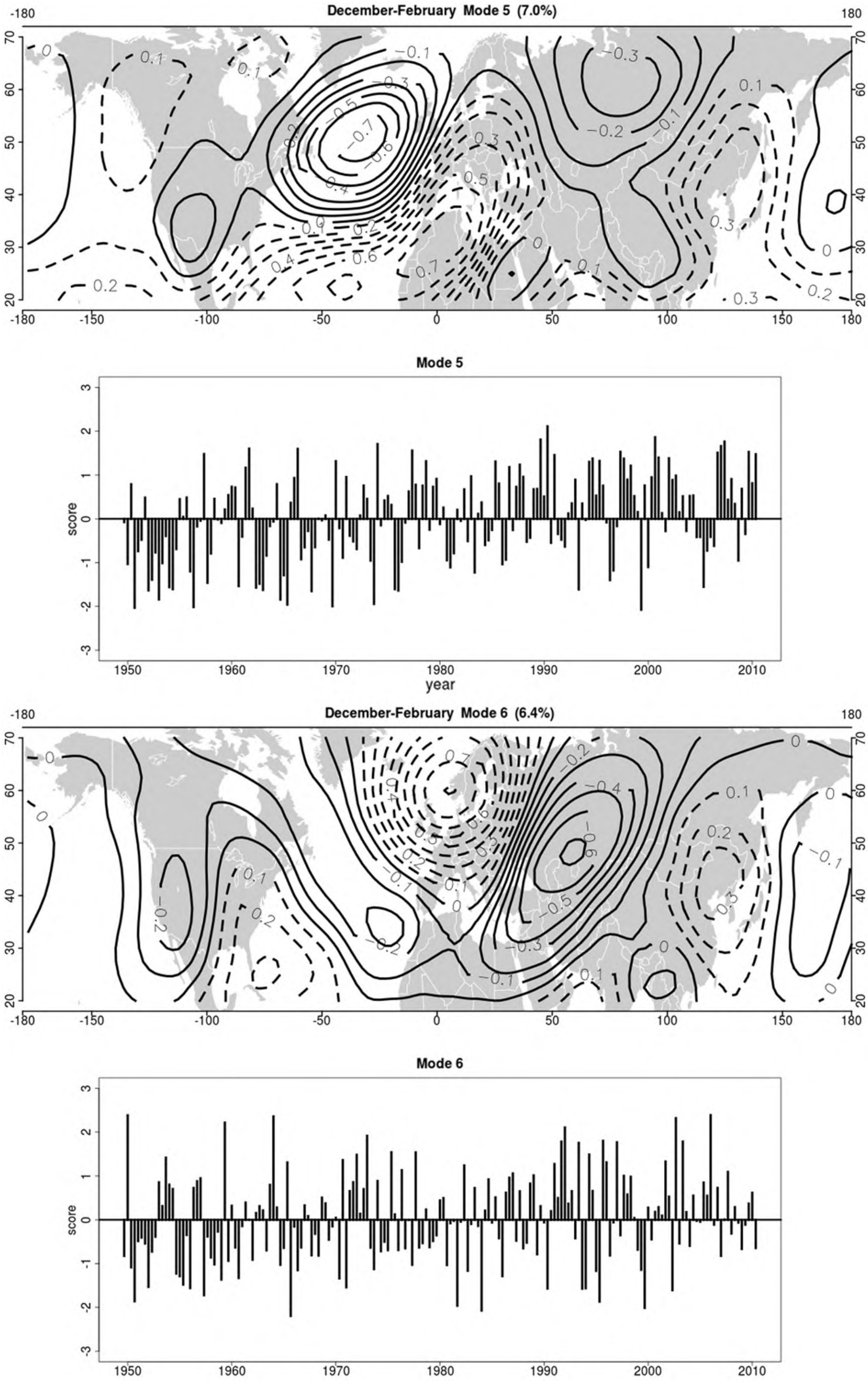


**Fig. 3.** East Atlantic pattern (upper figure), Eurasian Type 2 pattern (central figure), and Eurasian Type 1 pattern (lower figure), represented by the fifth, sixth, and seventh mode, respectively, from VARIMAX-rotated PCA of winter (December–February) area-weighted 500 hPa geopotential heights of the Northern Hemisphere between 20°N and 70°N. Time series show the corresponding monthly PC scores from December 1949 to February 2010. PCA is based on standardised monthly anomaly fields from the NCEP/NCAR reanalysis (Kalnay et al., 1996; Kistler et al., 2001). Ten PCs were kept with 73.2% of total variance explained. Numbers in parentheses in title: percent of variance explained by each mode.

with around 30% for winter temperatures and much less for the other seasons and for other climate parameters.

Related findings concerning the limited explanatory power of CTs for surface climate variations have been reported by Widmann and Schär (1997) and Goodess and Jones (2002) for precipitation during the second half of the 20th century in Switzerland and over the Iberian Peninsula, respectively, and by Philipp et al. (2007) and Parker (2009) with respect to temperature variations since the mid 19th century in Central Europe and Central England, respectively.

The reasons for the incapability of explaining long-term variations in surface climate by corresponding frequency changes of CTs are substantial non-stationarities in the relationship between CTs and surface

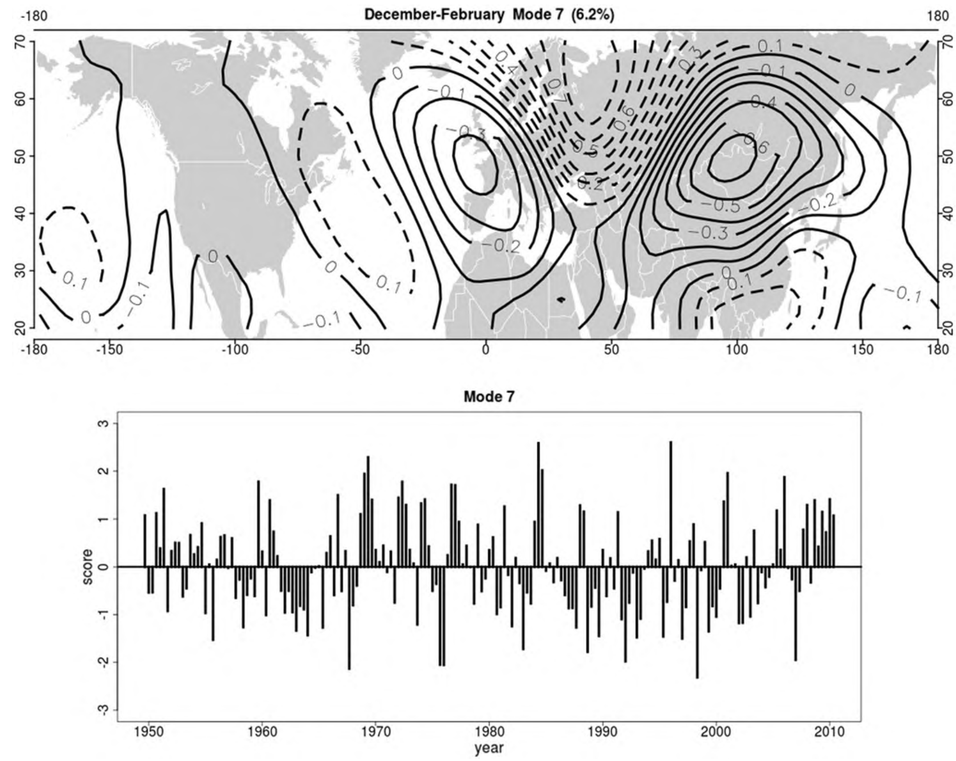


**Fig. 3** (*continued*).

climate—or so called within-type variations of CTs (Barry et al., 1981; Beck et al., 2007) denoting that CTs undergo temporal changes in their surface climate characteristics.

For winter precipitation 1961–1990 in Switzerland, Widmann and Schär (1997) found increasing trends in precipitation sums related to westerly, southerly and mixed CTs determined for the alpine region. Increasing trends in the probability of rainfall during winter (1958-1997) for the majority of objectively derived CTs have been documented for the Iberian Peninsula by Goodess and Jones (2002). Jones and Lister (2009) detected a slight within-type warming over Europe in the period 1971–2000 for several CTs in all seasons except autumn. Focusing on precipitation extremes in Central Europe in 1850–2003, Jacobeit et al. (2009) found a long-term within-type increase in the number of extreme precipitation events in winter for a westerly/northwesterly flow type. Concerning (positive) temperature extremes, within-type increases in the number of events appear for three zonal flow patterns in winter, whereas distinct multi-decadal variations turned out for an extended Azores high pattern in summer (Jacobeit et al., 2009).

Several more comprehensive analyses of within-type variations of CTs for the European region have been performed utilising reconstructed monthly sea level pressure data since the mid 17th century for the classification of CTs. Beside marked decadal to multi-decadal-scale variations, the most striking long-term changes in climatic within-type characteristics appear for several variants of westerly CTs with increasing trends in within-type temperatures and as well within-type precipitation in winter (Beck et al., 2001, 2007; Jacobeit et al., 2003; Küttel et al., 2010). For summer a long-term decrease in temperatures related to westerly CTs can be stated particularly since the late 18th century (Beck et al., 2001, 2007; Jacobeit et al., 2003). To determine the fractions of the observed variations in temperature and precipitation that may be attributed to the above mentioned within-type changes, Beck et al. (2007) and Küttel et al. (2010) applied a simple decomposition scheme (Perry and Barry, 1973). The main finding from these studies is that large parts (between around 40% and more than 80% depending on season, region and variable) of the temperature and precipitation variations since the mid 18th century can be ascribed to within-type changes of CTs. Estimating the importance of within-type variations for moving 31-year periods from 1780 to 1995 reveals the existence of distinct variations on decadal and multi-decadal time-scales rather than any clear-cut long-term trends (Beck et al., 2007).

The documented within-type variations of CTs reflect non-stationarities in the relationships between CTs and surface climate. They may originate from either dynamical or climatic sources (e.g. Beck et al., 2007). Variations in dynamic properties of CTs include for instance changes in the pressure gradients between centers of action or changes in vorticity characteristics of CTs. Such internal dynamical variations have been reported for varying CTs stemming from different classifications. Most prominent in this context are changes in vorticity and the meridional pressure gradient within westerly CTs during winter that can partly be related to corresponding variations in temperature and precipitation connected to these CTs (Jacobeit et al., 2003; Beck et al., 2007; Küttel et al., 2010). Another example refers to circulation pattern sequences which are linked with prominent discharge events (daily events with distinctly increasing discharge exceeding specific thresholds) in Central Europe only in case of enhanced cyclonic vorticity (Jacobeit et al., 2006).

However, not all observed non-stationarities can be attributed to varying circulation properties within CTs, but may rather originate from variations in superordinate driving forces, like variations in the large-scale forcing from the atmosphere, land, or ocean and/ or due to external forcing factors. First to mention in this context are variations associated with enhanced radiative forcing due to anthropogenic greenhouse gas emissions. However, a systematic post-industrial increase in within-type variability of CTs that should be expected from a major influence of this potential factor is not discernible (Beck et al., 2007; Küttel et al., 2010). More likely than a direct influence of 20th century global warming on within-type temperatures of specific CTs are related indirect effects for instance by means of altered sea surface temperatures. A distinct influence of North Atlantic sea surface temperatures on climatic properties of CTs has been substantiated by Jones et al. (1999)

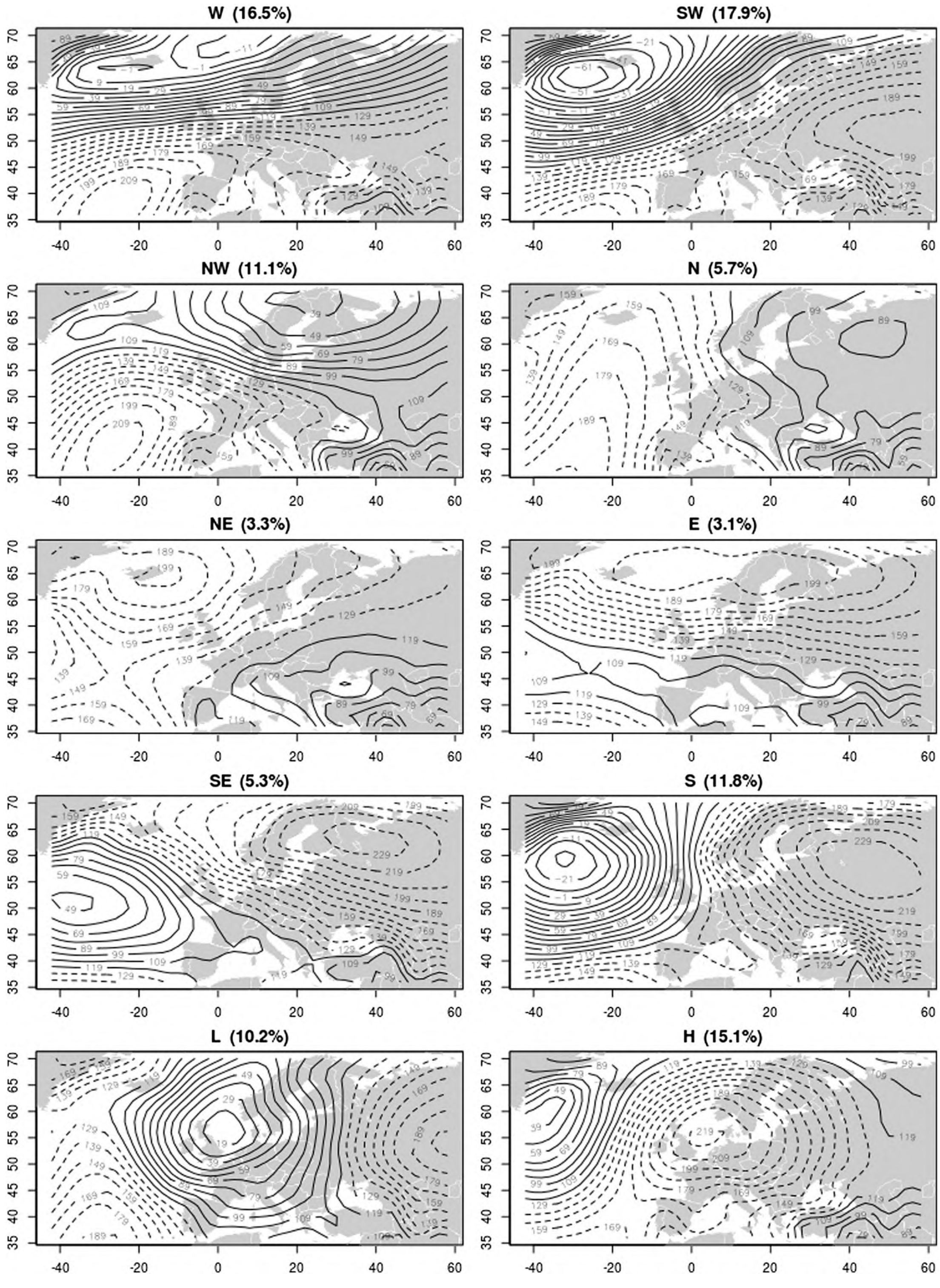


**Fig. 4.** 1000 hPa geopotential height (gpm) composites for 10 North Atlantic–European circulation types resulting from an objective circulation type classification (Beck et al., 2007) applied to daily 1000 hPa geopotential height fields between 42°W and 58°E and 36°N and 70°N for the period 1950–2010 from the NCEP/NCAR 20th century reanalysis data set (Compo et al., 2011). Circulation types reflect main directions of large-scale air flow (W, SW, NW, N, NE, E, SE, S) and central low or high pressure patterns (L, H). Numbers in brackets indicate relative occurrence frequencies of circulation types in the period 1950–2010.

and Parker (2009) for CTs inducing the advection of maritime air masses to Europe. Beside variations in sea surface temperatures, also changes in earth surface characteristics in general should be considered, as within-type changes have been ascertained as well for CTs related to the advection of air-masses from continental regions (Jacobeit et al., 2003; Beck et al., 2007; Küttel et al., 2010).

### 3.3. Mechanisms

Non-stationarities in the atmospheric circulation over the North Atlantic–European area and consequently in the relationships with regional climate are related to changes in the patterns' temporal behaviour, expressed by changes in frequency, amplitude, persistence,

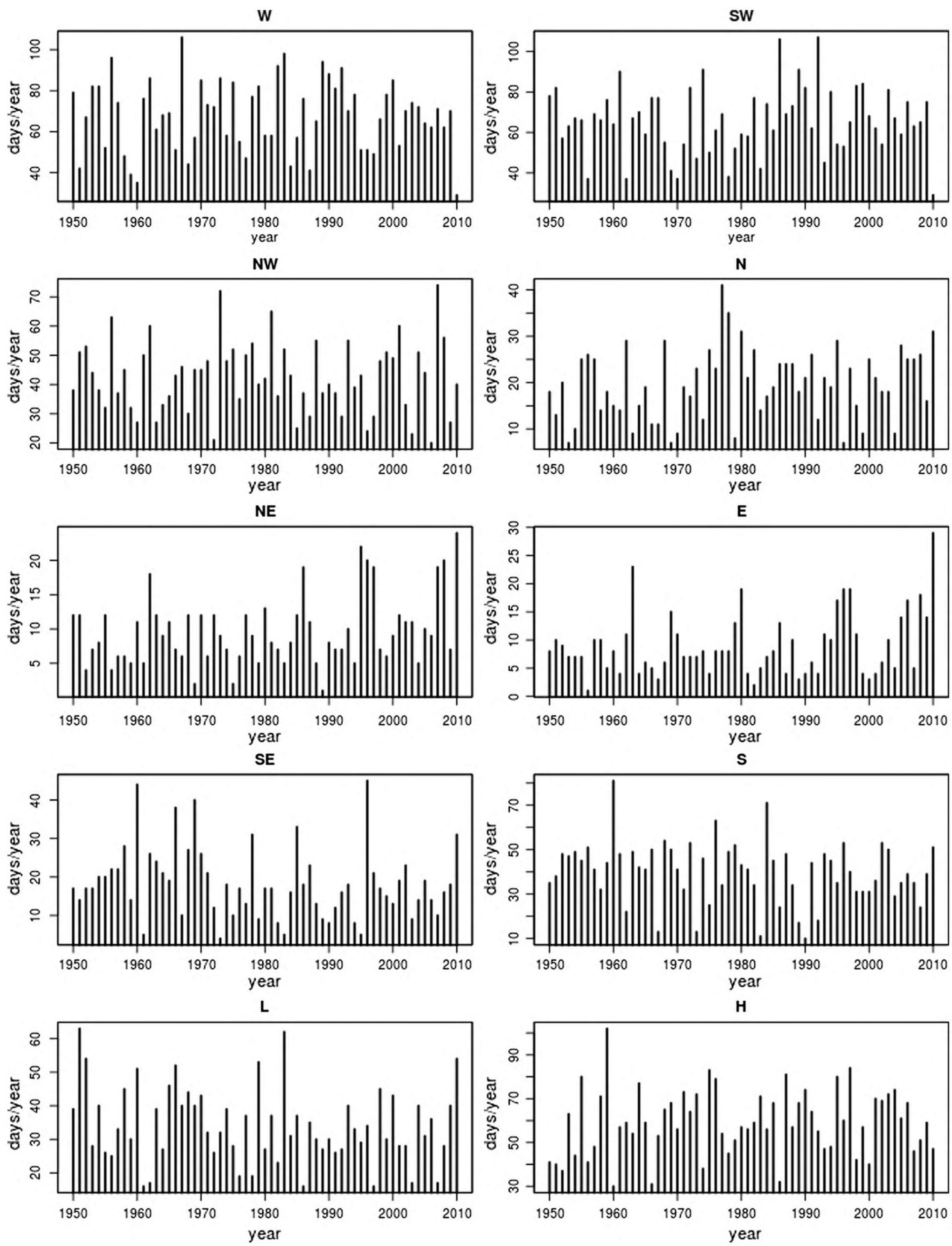


**Fig. 5.** Time series of occurrence frequencies (days/year) of the 10 North Atlantic–European circulation types from Fig. 4.

and succession of the patterns. Within the representation of the atmospheric circulation as large-scale modes of variability (Section 3.1), changes in the preferred phases of the patterns are also important. The commonly non-linear character of the phases has an important effect in this context. Non-stationarities can also be attributed to changes in the spatial structure of the patterns like changes of the intensity and spatial location of the centres of action of the large-scale modes and changes in the preferred flow patterns of CTs. Furthermore, within-type changes of CTs play an important role for the generation of non-stationarities. These within-type changes involve, for instance, changes in pressure gradients, vorticity parameters as well as type-specific temperature and humidity characteristics.

The large-scale state of the atmosphere can be defined by a set of meteorological variables which define a state space. The dynamical evolution of the climate system can be thought of as the motion of a point through this multi-dimensional state space (Stephenson et al., 2004). A climate pattern can be seen as a direction in state space and the associated index (e.g. principal component time series) is the projection onto this direction. Consequently, non-stationarities in patterns, evident in changes of the patterns temporal characteristics and spatial structure, can result from the chaotic nonlinear dynamical behaviour of the system. They are associated with the low-frequency variability of the system, but can also arise from high-order stochastic forcing.

Physical mechanisms underlying non-stationarities in the North Atlantic-European domain involve atmospheric internal variability, but also interactions of the atmosphere with the sea and land surfaces. The North Atlantic Ocean plays an important role in this context. Besides local interactions, teleconnections, e.g. with ENSO, can have a part in generating non-stationarities. External forcing has also to be taken into account, as discussed for instance in the scope of an impact of enhanced radiative forcing on CTs.

## 4. Discussion and conclusions

Non-stationarities in major modes of climate variability (ENSO, PNA, NAO) during the last century have been analysed. The analysis of the large-scale modes of climate variability was followed by a detailed analysis of non-stationarities with an explicitly regional focus on the North Atlantic–European area. Table 1 summarises the results by means of an overview of the non-stationarities and their causes and mechanisms for each major mode as well as for the minor modes and patterns.

Non-stationarities in climate variability are associated with different time scales. Both the atmosphere and oceans are involved in generating climate variability, and the corresponding time scales are quite different. Atmospheric processes become fully effective for all the climate variability on inter-annual to longer timescales, whereas oceanic processes will be fully effective only for climate variability at periods longer than their adjustment time. Thus, the oceans provide long-term memory, while the atmosphere provides stochastic forcing that can help to excite the low-frequency ocean fluctuations (Liu and Alexander, 2007). Even if climate variability generated within the coupled ocean–atmosphere system involves deterministic processes with a set time-scale, the complexity of processes as well as the additional random forcing make the determination of specific time scales of variability a complex task. This issue is further complicated by sea ice processes and land surface processes interacting with ocean–atmosphere variability.

If we observe large long-term deviations, the question arises whether this is just an expression of natural process dynamics around a fixed mean or whether this is part of fundamental structural changes. Non-stationarities can result from long-term excursions of an overall stationary process due to long-term persistence ("Hurst phenomenon", Hurst, 1951; Koutsoyiannis and Montanari, 2007; Lins and Cohn, 2011). Thus, non-stationarity by itself does not need to imply any change in the basic state of the system. So-called autoregressive integrated moving average processes generate non-stationary sequences for fixed parameters (Wunsch, 1999). From these considerations it becomes clear that non-stationarities can arise from fundamental climate change, as anticipated under further enhanced greenhouse gas forcing for instance, but that non-stationarities can also just be a part of natural climate variability. Strong inter-decadal to inter-centennial modulations can thereby be just a result of random forcing, as has been shown for ENSO behaviour by Wittenberg (2009). Non-stationarities in natural climate variability can be attributed to stochastic forcing due to high-frequency weather oscillations, to deterministic chaos arising from internal non-linearities of the natural modes of the atmosphere alone or the coupled atmosphere–ocean system, and to interactions between these modes. For instance, Pacific decadal variability is seen as a weakly damped coupled atmosphere–ocean oscillation. As summarised by Deser et al. (2012), basin-scale stochastic atmospheric forcing drives a delayed oceanic response via Rossby wave adjustment. The associated changes in lateral heat flux convergence along the Kuroshio–Oyashio Extension force SST anomalies that drive a weak atmospheric response needed to sustain the oscillation. Furthermore, the tropical Pacific plays a significant role by modifying and responding to the Pacific decadal variability.

Thus, non-stationarities can be an expression of non-linearities in the climate system. An example is the non-linearity of ENSO which resulted, according to Burgers and Stephenson (1999), in the predominance of El Niño events compared to La Niña during the second half of the 20th century. Besides, SST anomalies associated with ENSO can generate atmospheric waves that emanate into the extra-tropics. These waves form in preferred locations and result for example in the strengthening of the Aleutian centre of action in response to ENSO. However, the ENSO-driven extra-tropical circulation anomalies project not only on the PNA pattern, but on several different teleconnection patterns (Liu and Alexander, 2007). This induces differing changes in local climate.

Non-stationarities in the large-scale modes of atmospheric variability and consequently in the regional climate affected by these modes are associated with specific trends and low-frequency variability of the modes leading to a dominance of the positive or negative phases. In combination with non-linearity in the phases of the modes, non-stationarity can result. Another very important source for non-

**Table 1**
Overview of non-stationarities and their causes and mechanisms. For details see text.

| | Non-stationarities | Causes and mechanisms |
|---|---|---|
| **ENSO** | • decadal and multi-decadal changes in mean and variability<br>• "climate shift" in the mid-1970s<br>• changes in teleconnections | • non-linearity of ENSO<br>• stochastic processes within seasonal and inter-annual ENSO characteristics<br>• modulation of coupled atmosphere-ocean system<br>• multiple decadal climate modes<br>• changes of background state<br>• external forcing (volcanic, solar, anthropogenic greenhouse gases) |
| **PNA** | • multiple "regime shifts"<br>• changes in strength and spatial structure of pattern anomaly centres<br>• changes of PNA–climate relationships | • internal oscillations in coupled atmosphere–ocean system<br>• multiple modes of variability<br>• changes of oceanic forcing (ENSO, NPGO)<br>• long-term persistence<br>• external forcing (solar) |
| **NAO** | • variability of temporal modes<br>• changes in inter- to multidecadal variability<br>• changes in strength and spatial structure of anomaly centres<br>• changes of NAO–climate relationships | • non-linearity of NAO phases<br>• multiple NAO modes<br>• atmospheric low-frequency variability<br>• non-linearity dependence on background flow<br>• changes in relationships with stratosphere, ENSO, QBO<br>• long-range persistence<br>• external forcing (solar, volcanic, anthropogenic greenhouse gases) |
| **Non-NAO modes and CTs** | • changes in temporal behaviour of patterns<br>• changes in spatial structure of patterns<br>• within-type changes<br>• changes of circulation-climate relationships | • non-linearity of pattern phases<br>• atmospheric low-frequency variability<br>• high-order stochastic forcing<br>• interactions of atmosphere with land and sea surfaces<br>• external forcing (anthropogenic greenhouse gases) |

stationarities is the change of the intrinsic structure of atmospheric variability. This is associated with changes of the mean pattern states and/or with intensity changes and spatial shifts of pattern-specific centres of variation. Furthermore, changes in the overall importance of a pattern and changes in the relationships and interactions between the different patterns can become evident. Sea and land surface effects, remote influences (e.g. from ENSO) and external forcings (e.g. large tropical volcanic eruptions, anthropogenic warming) most likely contribute to the generation of non-stationarities.

Non-stationarities are usually dealt with by appropriate time series analysis techniques of observational records. In general, any observed non-stationarity is related to multiple processes which contribute in differing amounts. The decision on whether or not to include non-stationary relationships into further analyses depends strongly on the magnitude of the non-stationarity, its statistical evidence, and its impact. However, these considerations are not trivial. In case of large abrupt changes, a non-stationarity might be easily detected, but in case of slow changes of the state or the processes in a system, detection can be much more difficult.

Despite the potential drawback not to be able to determine all underlying physical mechanisms, a careful analysis of physical plausibility is essential. For this purpose statistical analysis tools as well as a hierarchy of simulation models of the atmospheric, oceanic and coupled ocean–atmosphere circulation as well as complex earth system models can be applied. The motivation for analysing the mechanisms underlying non-stationarities is manifold: (i) It serves to exclude the possibility that the found non-stationarities are just due to inhomogeneities in the analysed data. For instance, in the recent period one of the most relevant temporal inhomogeneity is the introduction of satellite observations into reanalysis data (e.g. Bengtsson et al., 2004). (ii) Statistical techniques are often based on the stationarity assumption. One example is water resources planning and design (Milly et al., 2008). Determination of the nature of observed non-stationarities can aid in the selection and development of appropriate tools and thus provides a basis to enhance confidence in the results. (iii) Non-stationarities can also arise within the context of climate change, and it is essential for the reliability of future projections to know the causes and mechanisms of possible non-stationarities. For instance, climate change associated with the greenhouse gas increases could substantially change the characteristics of ENSO (Collins et al., 2010) and thus the teleconnections from the tropics (Meehl et al., 2006). Zhou et al. (2014) show that the ENSO-forced part of the PNA moves eastward and is intensified under climate warming conditions, leading to intensified rainfall anomalies on the west coast of North America. Stevenson (2012) reports that in general circulation models which show ENSO amplitude increases with greenhouse gas increases, teleconnections tend to strengthen, but do not shift position, whereas in models with no ENSO amplitude increases, the teleconnections are weakened and shifted polewards and eastwards.

Certainly, non-stationarities can be caused by abrupt changes in the climate system due to external forcing factors, but it should be particularly emphasised that non-stationarities are also an intrinsic part of natural climate variability. Insufficient consideration of the stationarity issue in climatology or in related research fields can have severe consequences. One example is regional climate change assessments obtained by statistical downscaling. Usually a statistical model is calibrated in specific observational periods and subsequently validated in periods independent from the calibration. If significant non-stationarities occur between these periods, the model will fail to capture the changed conditions (e.g. Hertig and Jacobeit, 2013). This issue gains even further importance, if significant non-stationarities occur in the projection period under enhanced greenhouse gas forcing. As for non-stationarities in the observational record, we can expect that non-stationarities will also be part of future climate variability (e.g. Ullmann et al., 2013; Hertig and Jacobeit, 2014a,b). Non-stationarities might even be enhanced due to the interaction of natural variability and anthropogenic forcing factors like further increased greenhouse gas concentrations and land use changes.

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## References

Álvarez-García, F.J., OrtizBevia, M.J., CabosNarvaez, W.D., 2011. On the structure and teleconnections of North Atlantic decadal variability. J. Clim. 24, 2209–2223.

Ambaum, M.H.P., Hoskins, B.J., Stephenson, D.B., 2001. Arctic Oscillation or North Atlantic Oscillation? J. Clim. 14, 3495–3507.

Ashok, K., Behera, S.K., Rao, S.A., Weng, H., Yamagata, T., 2007. El Niño Modoki and its possible teleconnection. J. Geophys. Res. Oceans 112, C11007. http://dx.doi.org/10.1029/2006JC003798.

Barnston, A.G., Livezey, R.E., 1987. Classification, seasonality and persistence of low-frequency atmospheric circulation patterns. Mon. Weather Rev. 115, 1083–1126.

Barry, R.G., Kiladis, G., Bradley, R.S., 1981. Synoptic climatology of the Western United States in relation to climatic fluctuations during the twentieth century. J. Clim. 1, 97–113.

Battisti, D.S., Hirst, A.C., 1989. Interannual variability in the tropical atmosphere-ocean system: influence of the basic state and ocean geometry. J. Atmos. Sci. 46, 1687–1712.

Beck, C., Jacobeit, J., Philipp, A., 2001. Variability of North-Atlantic–European Circulation Patterns since 1780 and Corresponding Variations in Central European Climate. In: Brunet India, M., Lopez Bonillo, D. (Eds.), Detecting and Modelling Regional Climate Change. Springer Verlag, Berlin Heidelberg New York, pp. 321–332.

Beck, C., Jacobeit, J., Jones, P.D., 2007. Frequency and within-type variations of large scale circulation types and their effects on low-frequency climate variability in Central Europe since 1780. Int. J. Climatol. 27, 473–491.

Bengtsson, L., Hodges, K.I., Hagemann, S., 2004. Sensitivity of the ERA40 reanalysis to the observing system: determination of the global atmospheric circulation from reduced observations. Tellus A 56, 456–471.

Beranová, R., Huth, R., 2008. Time variations of the effects of circulation variability modes on European temperature and precipitation in winter. Int. J. Climatol. 28, 139–158.

Blackmon, M.L., Geisler, J.E., Pitcher, E.J., 1983. A general circulation model study of January climate anomaly patterns associated with interannual variation of equatorial Pacific sea surface temperatures. J. Atmos. Sci. 40, 1410–1425.

Bond, N.A., Overland, J.E., Spillane, M., Stabeno, P., 2003. Recent shifts in the state of the North Pacific. Geophys. Res. Lett. 30. http://dx.doi.org/10.1029/2003GL018597.

Brandefelt, J., 2006. Atmospheric modes of variability in a changing climate. J. Clim. 19, 5934–5943.

Brönnimann, S., Xoplaki, E., Casty, C., Pauling, A., Luterbacher, J., 2007. ENSO influence on Europe during the last centuries. Clim. Dyn. 28, 181–197.

Budikova, D., 2012. Northern Hemisphere climate variability: character, forcing mechanisms, and significance of the North Atlantic/Arctic Oscillation. Geogr. Compass 1–22.

Burgers, G., Stephenson, D.B., 1999. The "normality" of El Niño. Geophys. Res. Lett. 26, 1027–1030.

Cahynová, M., Huth, R., 2009. Changes of atmospheric circulation in central Europe and their influence on climatic trends in the Czech Republic. Theor. Appl. Climatol. 96, 57–68.

Cahynova, M., Huth, R., 2010. Circulation vs. climatic changes over the Czech Republic: a comprehensive study based on the COST733 database of atmospheric circulation classifications. Phys. Chem. Earth 35, 422–428.

Cassou, C., Terray, L., Hurrell, J., Deser, C., 2004. North Atlantic winter climate regimes: spatial asymmetry, stationarity with time, and oceanic forcing. J. Clim. 17, 1055–1068.

Cayan, D.R., 1992. Latent and sensible heat flux anomalies over the Northern Oceans: the connection to monthly atmospheric circulation. J. Clim. 5, 354–369.

Chen, D., Hellstrom, C., 1999. The influence of the North Atlantic Oscillation on the regional temperature variability in Sweden: spatial and temporal variations. Tellus 51A, 505–516.

Chowdary, J.S., Xie, S.-P., Tokinaga, H., Okumura, Y.M., Kubota, H., Johnson, N., Zheng, X.-T., 2012. Interdecadal variations in ENSO teleconnections to the Indo-Western Pacific for 1870-2007. J. Clim. 25, 1722–1744.

Christensen, J.H., Krishna Kumar, K., Aldrian, E., An, S.-I., Cavalcanti, I.F.A., de Castro, M., Dong, W., Goswami, P., Hall, A., Kanyanga, J.K., Kitoh, A., Kossin, J., Lau, N.-C., Renwick, J., Stephenson, D.B., Xie, S.-P., Zhou, T., 2013. Climate Phenomena and Their Relevance for Future Regional Climate Change. In: Stocker, T.F., Qin, D., Plattner, G.-K., Tignor, M., Allen, S.K., Boschung, J., Nauels, A., Xia, Y., Bex, V., Midgley, P.M. (Eds.), Climate Change 2013: The Physical Science Basis. Contribution of Working Group I to the Fifth Assessment Report of the Intergovernmental Panel on Climate Change. Cambridge University Press, Cambridge, United Kingdom and New York, NY, USA.

Collins, M., An, S.-I., Cai, W., Ganachaud, A., Guilyardi, E., Jin, F.-F., Jochum, M., Lengaigne, M., Power, S., Timmermann, A., Vecchi, G., Wittenberg, A., 2010. The impact of global warming on the tropical Pacific Ocean and El Niño. Nat. Geosci. 3, 391–397.

Compo, G.P., Whitaker, J.S., Sardeshmukh, P.D., Matsui, N., Allan, R.J., Yin, X., Gleason, B.E., Vose, R.S., Rutledge, G., Bessemoulin, P., Brönnimann, S., Brunet, M., Crouthamel, R.I., Grant, A.N., Groisman, P.Y., Jones, P.D., Kruk, M., Kruger, A.C., Marshall, G.J., Maugeri,

M., Mok, H.Y., Nordli, Ø., Ross, T.F., Trigo, R.M., Wang, X.L., Woodruff, S.D., Worley, S.J., 2011. The twentieth century reanalysis project. Q. J. R. Meteorolog. Soc. 137, 1–28.
Czaja, A., Frankignoul, C., 2002. Observed impact of Atlantic SST anomalies on the North Atlantic Oscillation. J. Clim. 15, 606–623.
Deser, C., Phillips, A.S., Hurrell, J.W., 2004. Pacific Interdecadal Climate Variability: linkages between the Tropics and the North Pacific during Boral Winter since 1900. J. Clim. 17, 3109–3124.
Deser, C., Phillips, A.S., Tomas, R.A., Okumura, Y.M., Alexander, M.A., Capotondi, A., Scott, J.D., Kwon, Y.-O., Ohba, M., 2012. ENSO and Pacific decadal variability in the community climate system model version 4. J. Clim. 25, 2622–2651.
Di Lorenzo, E., Schneider, N., Cobb, K.M., Franks, P.J.S., Chhak, K., Miller, A.J., McWilliams, J.C., Bograd, S.J., Arango, H., Curchitser, E., Powell, T.M., Rivière, P., 2008. North Pacific Gyre Oscillation links ocean climate and ecosystem change. Geophys. Res. Lett. 35, L08607. http://dx.doi.org/10.1029/2007GL032838.
Diao, Y., Xie, S.-P., Luo, D., 2014. Asymmetry of winter European surface air temperature extremes and the North Atlantic Oscillation. J. Clim. http://dx.doi.org/10.1175/JCLI-D-13-00642.1.
Diaz, H.F., Hoerling, M.P., Eischeid, J.K., 2001. ENSO variability, teleconnections and climate change. Int. J. Climatol. 21, 1845–1862.
Dijkstra, H.A., 2006. The ENSO phenomenon: theory and mechanisms. Advances in Geosciences 6. European Geosciences Union (EGU), pp. 3–15.
Dong, B., Sutton, R.T., Woollings, T., 2011. Changes of interannual NAO variability in response to greenhouse gas forcing. Clim. Dyn. 37, 1621–1641.
Espinoza Villar, J.C., Ronchail, J., Guyot, J.L., Cochonneau, G., Naziano, F., Lavado, W., De Oliveira, E., Pombosa, R., Vauchel, P., 2009. Spatio-temporal rainfall variability in the Amazon basin countries (Brazil, Peru, Bolivia, Colombia, and Ecuador). Int. J. Climatol. 29, 1574–1594.
Feldstein, S.B., 2000. The timescale, power spectra, and climate noise properties of teleconnection patterns. J. Clim. 13, 4430–4440.
Franzke, Ch., Feldstein, S.B., Lee, S., 2011. Synoptic analysis of the Pacific–North American teleconnection pattern. Q. J. R. Meteorol. Soc. 137, 329–346.
Gastineau, G., D'Andrea, F., Frankignoul, C., 2013. Atmospheric response to the North Atlantic Ocean variability on seasonal to decadal time scales. Clim. Dyn. 40, 2311–2330.
Gillett, N.P., Fyfe, J.C., 2013. Annular mode changes in the CMIP5 simulations. Geophys. Res. Lett. 40, 1189–1193.
Gimeno, L., de la Torre, L., Nieto, R., García, R., Hernández, E., Ribera, P., 2003. Changes in the relationship NAO-Northern hemisphere temperature due to solar activity. Earth Planet. Sci. Lett. 206, 15–20.
Goodess, C.M., Jones, P.D., 2002. Links between circulation and changes in the characteristics of Iberian rainfall. Int. J. Climatol. 22, 1593–1615.
Goodkin, N.F., Hughen, K.A., Donley, S.C., Curry, W.B., 2008. Increased multidecadal variability of the North Atlantic Oscillation since 1781. Nat. Geosci. 2008, 844–848.
Graham, N.E., 1994. Decadal scale variability in the 1970s and 1980s: observations and model results. Clim. Dyn. 10, 60–70.
Greatbatch, R.J., Lu, J., Peterson, K.A., 2004. Nonstationary impact of ENSO on Euro-Atlantic winter climate. Geophys. Res. Lett. 31, L02208. http://dx.doi.org/10.1029/2003GL018542.
Halpert, M.S., Ropelewski, C.F., 1992. Surface temperature patterns associated with the Southern Oscillation. J. Clim. 5, 577–593.
Hanna, E., Cropper, T.E., Jones, P.D., Scaife, A.A., Allan, R., 2014. Recent seasonal asymmetric changes in the NAO (a marked summer decline and increased winter variability) and associated changes in the AO and Greenland Blocking Index. Int. J. Climatol. http://dx.doi.org/10.1002/joc.4157.
Hare, S.R., Mantua, N.J., 2000. Empirical evidence for North Pacific regime shifts in 1977 and 1989. Prog. Oceanogr. 47, 103–145.
Haylock, M.R., Peterson, T.C., Alves, L.M., Ambrizzi, T., Anunciação, Y.M.T., Baez, J., Barros, V.R., Vincent, L.A., 2006. Trends in total and extreme South American rainfall in 1960-2000 and links with sea surface temperature. J. Clim. 19, 1490–1512.
Hertig, E., Jacobeit, J., 2013. A novel approach to statistical downscaling considering nonstationarities: application to daily precipitation in the Mediterranean area. J. Geophys. Res. Atmos. 118, 520–533.
Hertig, E., Jacobeit, J., 2014a. Variability of weather regimes in the North Atlantic–European area: past and future. Atmos. Sci. Lett. http://dx.doi.org/10.1002/asl2.505.
Hertig, E., Jacobeit, J., 2014b. Considering observed and future non-stationarities in statistical downscaling of Mediterranean precipitation. Theor. Appl. Climatol. http://dx.doi.org/10.1007/s00704-014-1314-9.
Higgins, R.W., Mo, K.C., 1997. Persistent North Pacific circulation anomalies and the tropical intraseasonal oscillation. J. Clim. 10, 223–244.
Hurrell, J., Deser, C., 2009. North Atlantic climate variability: the role of the North Atlantic Oscillation. J. Mar. Syst. 78, 28–41.
Hurrell, J., van Loon, H., 1997. Decadal variations in climate associated with the North Atlantic Oscillation. Clim. Chang. 36, 301–326.
Hurrell, J., Kushnir, Y., Ottersen, G., Visbeck, M. (Eds.), 2003. The North Atlantic Oscillation: climatic significance and environmental impact. Geophys. Monogr. Ser. 134. AGU, Washington, D. C..
Hurst, H.E., 1951. Long term storage capacities of reservoirs. Trans. Am. Soc. Civ. Eng. 116, 776–806.
Huth, R., 2001. Disaggregating climatic trends by classification of circulation patterns. Int. J. Climatol. 21, 135–153.
Huth, R., Beck, C., Philipp, A., Demuzere, M., Untstrul, Z., Cahynova, M., Kysely, J., Tveito, O.E., 2008. Classifications of atmospheric circulation patterns: recent advances and applications. Ann. N.Y. Acad. Sci. 1146, 105–152.
Jacobeit, J., Jönsson, P., Bärring, L., Beck, Ch., Ekström, M., 2001. Zonal indices for Europe 1780–1995 and running correlations with temperature. Clim. Chang. 48, 219–241.
Jacobeit, J., Wanner, H., Luterbacher, J., Beck, C., Philipp, A., Sturm, K., 2003. Atmospheric circulation variability in the North-Atlantic–European area since the mid-seventeenth century. Clim. Dyn. 20, 341–352.
Jacobeit, J., Philipp, A., Nonnenmacher, M., 2006. Atmospheric circulation dynamics linked with prominent discharge events in Central Europe. Hydrol. Sci. J. 51, 946–965.
Jacobeit, J., Rathmann, J., Philipp, A., Jones, P.D., 2009. Central European precipitation and temperature extremes in relation to large-scale atmospheric circulation types. Meteorol. Z. 18, 397–410.
Johnson, N.C., Feldstein, S.B., 2010. The Continuum of North Pacific sea level pressure patterns: intraseasonal, interannual, and interdecadal variability. J. Clim. 23, 851–867.
Jones, P.D., Lister, D.H., 2009. The influence of the circulation on surface temperature and precipitation patterns over Europe. Clim. Past 5, 259–267.
Jones, P.D., Jónsson, T., Wheeler, D., 1997. Extension to the North Atlantic Oscillation using early instrumental pressure observations from Gibraltar and south-west Iceland. Int. J. Climatol. 17, 1433–1450.
Jones, P.D., Horton, E.B., Folland, C.K., Hulme, M., Parker, D.E., Basnett, T.A., 1999. The use of indices to identify changes in climatic extremes. Clim. Chang. 42, 131–149.
Jung, Th., Hilmer, M., Ruprecht, E., Kleppek, S., Gulev, S., Zolina, O., 2003. Characteristics of the recent eastward shift of interannual NAO variability. J. Clim. 16, 3371–3381.
Kalnay, E., Kanamitsu, M., Kistler, R., Collins, W., Deaven, D., Gandin, L., Iredell, M., Saha, S., White, G., Woollen, J., Zhu, Y., Chelliah, M., Ebisuzaki, W., Higgins, W., Janowiak, J., Mo, K.C., Ropelewski, C., Wang, J., Leetmaa, A., Reynolds, R., Jenne, R., Joseph, D., 1996. The NCEP/NCAR 40-year reanalysis project. Bull. Am. Meteorol. Soc. 77, 437–471.
Kang, I.-S., No, H., Kucharski, F., 2014. ENSO amplitude modulation associated with the mean SST changes in the tropical central Pacific induced by Atlantic Multidecadal Oscillation. J. Clim. 27, 7911–7920.
Kenyon, J., Hegerl, G.C., 2008. Influence of modes of climate variability on global temperature extremes. J. Clim. 21, 3872–3889.
Kistler, R., Kalnay, E., Collins, W., Saha, S., White, G., Woollen, J., Chelliah, M., Ebisuzaki, W., Kanamitsu, M., Kousky, V., van den Dool, H., Jenne, R., Fiorino, M., 2001. The NCEP/NCAR 50-year reanalysis: monthly means CD-ROM and documentation. Bull. Am. Meteorol. Soc. 82, 247–268.
Knight, J.R., Folland, C.K., Scaife, A.A., 2006. Climate impacts of the Atlantic Multidecadal Oscillation. Geophys. Res. Lett. 33, L17706.
Kosaka, Y., Xie, S.-P., Lau, N.-C., Vecchi, G.A., 2013. Origin of seasonal predictability for summer climate over the Northwestern Pacific. PNAS 110, 7574–7579.
Koutsoyiannis, D., Montanari, A., 2007. Statistical analysis of hydroclimatic time series: uncertainty and insights. Water Resour. Res. 43, W05429.
Krichak, S.O., Alpert, P., 2005. Decadal trends in the East Atlantic–West Russia pattern and Mediterranean precipitation. Int. J. Climatol. 25, 183–192.
Kutiel, H., Maheras, P., Türkeş, M., Paz, S., 2002. North Sea–Caspian Pattern (NCP)–an upper level atmospheric teleconnection affecting the eastern Mediterranean—implications on the regional climate. Theor. Appl. Climatol. 72, 173–192.
Küttel, M., Luterbacher, J., Wanner, H., 2010. Multidecadal changes in winter circulation-climate relationship in Europe: frequency variations, within-type modifications, and long-term trends. Clim. Dyn. http://dx.doi.org/10.1007/s00382-009-0737-y.
Larkin, N.K., Harrison, D.E., 2005. Global seasonal temperature and precipitation anomalies during El Niño autumn and winter. Geophys. Res. Lett. 32. http://dx.doi.org/10.1029/2005GL022860.
Lau, N.-C., Nath, M.J., 2003. Atmosphere-ocean variations in the Indo-Pacific sector during ENSO episodes. J. Clim. 16, 3–20.
Leathers, D.J., Yarnal, B., Palecki, M.A., 1991. The Pacific/North American teleconnection pattern and United States Climate. Part I: regional temperature and precipitation Associations. J. Clim. 4, 517–528.
Li, J., Xie, S.-P., Cook, E.R., Morales, M.S., Christie, D.A., Johnson, N.C., Chen, F., D'Arrigo, R., Fowler, A.M., Gou, X., Fang, K., 2013. El Niño modulations over the past seven centuries. Nat. Clim. Chang. 3, 822–826.
Lins, H.F., Cohn, T.A., 2011. Stationarity: wanted dead or alive? J. Am. Water Resour. Assoc. 47, 475–480.
Liu, Z., 2012. Dynamics if interdecadal climate variability: a historical perspective. J. Clim. 25, 1963–1995.
Liu, Z., Alexander, M., 2007. Atmospheric bridge, oceanic tunnel, and global climatic teleconnections. Rev. Geophys. 45, RG2005. http://dx.doi.org/10.1029/2005RG000172.
Lorenz, E.N., 1951. Seasonal and irregular variations of the northern hemisphere sea-level pressure profile. J. Meteorol. 8, 52–59.
Luo, D., Gong, T., 2006. A possible mechanism for the eastward shift of interannual NAO action centers in last three decades. Geophys. Res. Lett. 33. http://dx.doi.org/10.1029/2006GL027860.
Mächel, H., Kapala, A., Flohn, H., 1998. Behaviour of the centres of action above the Atlantic since. Part I: characteristics of seasonal and interannual variability. Int. J. Climatol. 18, 1–22.
Mantua, N.J., Hare, S.R., Zhang, Y., Wallace, J.M., Francis, R.C., 1997. A Pacific interdecadal climate oscillation with impacts on salmon production. Bull. Am. Meteorol. Soc. 78, 1069–1079.
Mariotti, A., Zeng, N., Lau, K.-M., 2002. Euro-Mediterranean rainfall and ENSO—a seasonally varying relationship. Geophys. Res. Lett. 29, 59-1–59-4.
Meehl, G.A., Teng, H., Branstator, G., 2006. Future changes of El Niño in two global coupled climate models. Clim. Dyn. 26, 549–566.
Messié, M., Chavez, F., 2011. Global modes of sea surface temperature variability in relation to regional climate indices. J. Clim. 24, 4314–4331.
Michelangeli, P.A., Vautard, R., Legras, B., 1995. Weather regimes: recurrence and quasi stationarity. J. Atmos. Sci. 52, 1237–1256.
Miller, A.J., Cayan, D.R., Barnett, T.P., Graham, N.E., Oberhuber, J.M., 1994. The 1976–77 climate shift of the Pacific Ocean. Oceanography 7, 21–26.

Mills, T.C., 2004. Is the North Atlantic Oscillation a random walk? A comment with further results. Int. J. Climatol. 24, 377–383.

Milly, P.C.D., Betancourt, J., Falkenmark, M., Hirsch, R.M., Kundzewicz, Z.W., Lettenmaier, D.P., Stouffer, R.J., 2008. Stationarity is dead: whither water management? Science 319, 573–574.

Minobe, S., 1997. A 50–70 year climatic oscillation over the North Pacific and North America. Geophys. Res. Lett. 24, 683–686.

Mitchell, T.P., Wallace, J.M., 1992. The annual cycle in equatorial convection and sea surface temperature. J. Clim. 5, 1140–1156.

Neelin, J.D., Battisti, D.S., Hirst, A.C., Jin, F.F., Wakata, Y., Yamagata, T., Zebiak, S.E., 1998. ENSO theory. J. Geophys. Res 103 (C7), 14261–14290. http://dx.doi.org/10.1029/97JC03424.

Osborn, T.J., 2011. Winter 2009/2010 temperatures and a record-breaking North Atlantic Oscillation index. Weather 66, 19–21.

Overland, J., Rodionov, S., Minobe, S., Bond, N., 2008. North Pacific regime shifts: definitions, issues and recent transitions. Prog. Oceanogr. 77, 92–102.

Panagiotopoulos, F., Shahgedanova, M., Stephenson, D.B., 2002. A review of Northern Hemisphere winter-time tele connection patterns. J. Phys. IV France 12. http://dx.doi.org/10.1051/jp4:20020450.

Parker, D.E., 2009. Anomalies of central England temperature classified by air source. J. Clim. 22, 1069–1081.

Pauling, A., Luterbacher, J., Casty, C., Wanner, H., 2006. Five hundred years of gridded high-resolution precipitation reconstructions over Europe and the connection to large-scale circulation. Clim. Dyn. 26, 387–405.

Perry, A.H., Barry, R.G., 1973. Recent temperature changes due to changes in the frequency and average temperature of weather types over the British Isles. Meteorol. Mag. 102, 73–82.

Pezzulli, S., Stephenson, D.B., Hannachi, A., 2005. The variability of seasonality. J. Clim. 18, 71–88.

Philander, S.G., 1990. El Nino, La Nina and the Southern Oscillation. Int. Geophys. Ser. 46, 1990.

Philipp, A., Della-Marta, P.M., Jacobeit, J., Fereday, D.R., Jones, P.D., Moberg, A., Wanner, H., 2007. Long-term variability of daily North Atlantic–European pressure patterns since 1850 classified by simulated annealing clustering. J. Clim. 20, 4065–4095.

Philipp, A., Bartholy, J., Beck, C., Erpicum, M., Esteban, P., Fettweis, X., Huth, R., James, P., Jourdain, S., Kreienkamp, F., Krennert, T., Lykoudis, S., Michalides, S., Pianko-Kluczynska, K., Post, P., Rassilla Álvarez, D., Schiemann, R., Spekat, A., Tymvios, F.S., 2010. COST733CAT—a database of weather and circulation type classifications. Phys. Chem. Earth 35, 360–373.

Power, S.B., Smith, I.N., 2007. Weakening of the Walker Circulation and apparent dominance of El Niño both reach record levels, but has ENSO really changed? Geophys. Res. Lett. 34, L18702. http://dx.doi.org/10.1029/2007GL030854.

Pozo-Vázquez, D., Esteban-Parra, M.J., Rodrigo, F.S., Castro-Díez, Y., 2001. A study of NAO variability and its possible non-linear influences on European surface temperature. Clim. Dyn. 17, 701–715.

Qian, C., Wu, Z., Fu, C., Wang, D., 2011. On changing El Niño: a view from time-varying annual cycle, interannual variability, and mean state. J. Clim. 24, 6486–6500.

Rasmusson, E.M., Carpenter, T.H., 1982. Variations in tropical sea surface temperature and surface wind fields associated with the Southern Oscillation/El Niño. Mon. Weather Rev. 110, 354–384.

Rasmusson, E.M., Wallace, J.M., 1983. Meteorological aspects of the El Niño/Southern Oscillation. Science 222, 1195–1202.

Reid, P., Jones, P.D., Brown, O., Goodess, C.M., Davies, T.D., 2001. Assessments of the reliability of NCEP circulation data and relationships with surface climate data by direct comparisons with station based temperature, precipitation and mean sea level pressure data. Clim. Res. 17, 247–261.

Rodionov, S.N., 2004. A sequential algorithm for testing climate regime shifts. Geophys. Res. Lett. 31, L09204.

Rodionov, S.N., 2006. Use of prewhitening in climate regime shift detection. Geophys. Res. Lett. 33, L12707.

Ropelewski, C.F., Halpert, M.S., 1987. Global and regional scale precipitation patterns associated with the El Niño/Southern Oscillation. Mon. Weather Rev. 115, 1606–1626.

Ropelewski, C.F., Halpert, M.S., 1996. Quantifying Southern Oscillation-precipitaiton relationships. J. Clim. 9, 1043–1059.

Saji, N.H., Goswami, B.N., Vinayachandran, P.N., Yamagata, T., 1999. A dipole mode in the tropical Indian Ocean. Nature 401, 360–363.

Schlesinger, M.E., Ramankutty, N., 1994. An oscillation in the global climate system of period 65–70 years. Nature 367, 723–726.

Slonosky, V.C., Yiou, P., 2001. The North Atlantic Oscillation and its relationship with near surface temperature. Geophys. Res. Lett. 28, 807–810.

Slonosky, V.C., Jones, P.D., Davies, T.D., 2000. Variability of the surface atmospheric circulation over Europe, 1774–1995. Int. J. Climatol. 20, 1875–1897.

Slonosky, V.C., Jones, P.D., Davies, T.D., 2001. Atmospheric circulation and surface temperature in Europe from the 18th century to 1995. Int. J. Climatol. 21, 63–75.

Smith, T.M., Reynolds, R.W., Peterson, T.C., Lawrimore, J., 2008. Improvements to NOAA's Historical Merged Land–Ocean Surface Temperature Analysis (1880–2006). J. Clim. 21, 2283–2296.

Stephenson, D.B., Pavan, V., Bojariu, R., 2000. Is the North Atlantic Oscillation a random walk? Int. J. Climatol. 20, 1–18.

Stephenson, D.B., Hannachi, A., O'Neill, A., 2004. On the existence of multiple climate regimes. Q. J. R. Meteorol. Soc. 130, 583–605.

Stevenson, S.L., 2012. Significant changes to ENSO strength and impacts in the twenty-first century: Results from CMIP5. Geophys. Res. Lett. 39, L17703. http://dx.doi.org/10.1029/2012GL052759.

Suppiah, R., 2004. Trends in the southern oscillation phenomenon and Australian rainfall and changes in their relationship. Int. J. Climatol. 24, 269–290.

Thompson, D.W.J., Wallace, J.M., 1998. The Arctic Oscillation signature in the wintertime geopotential height and temperature fields. Geophys. Res. Lett. 25, 1297–1300.

Toniazzo, T., Scaife, A.A., 2006. The influence of ENSO on winter North Atlantic climate. Geophys. Res. Lett. 33, L24704.

Touchan, R., Xoplaki, E., Funkhouser, G., Luterbacher, J., Hughes, M.K., Erkan, N., Akkemik, Ü., Stephan, J., 2005. Reconstructions of spring/summer precipitation for the Eastern Mediterranean from tree-ring widths and its connection to large-scale atmospheric circulation. Clim. Dyn. 25, 75–98.

Trenberth, K.E., 1990. Recent observed interdecadal climate changes in the northern hemisphere. BAMS 71, 988–993.

Trenberth, K.E., Caron, J.M., 2000. The Southern Oscillation revisited: sea level pressures, surface temperatures, and precipitation. J. Clim. 13, 4358–4365.

Trenberth, K.E., Hoar, T.J., 1996. The 1990-1995 El Niño Southern Oscillation event: longest on record. Geophys. Res. Lett. 23, 57–60.

Trenberth, K.E., Hurrell, J.W., 1994. Decadal atmosphere-ocean variations in the Pacific. Clim. Dyn. 9, 303–319.

Trenberth, K.E., Shea, D.J., 1987. On the evolution of the Southern Oscillation. Mon. Weather Rev. 115, 3078–3096.

Ullmann, A., Fontaine, B., Roucou, P., 2013. Euro-Atlantic weather regimes and Mediterranean rainfall patterns: present-day variability and expected changes under CMIP5 projections. Int. J. Climatol. http://dx.doi.org/10.1002/joc.3864.

Van Oldenborgh, G.J., Burgers, G., 2005. Searching for decadal variations in ENSO precipitation teleconnections. Geophys. Res. Lett. 32, L15701. http://dx.doi.org/10.1029/2005GL023110.

Vicente-Serrano, S.M., López-Moreno, J.I., 2008. Nonstationary influence of the North Atlantic Oscillation on European precipitation. J. Geophys. Res. Atmos. 113, D20120.

Von Storch, H., Zwiers, F.W., 1999. Statistical Analysis in Climate Research. Cambridge University Press (495 pp.).

Wallace, J.M., Gutzler, D.S., 1981. Teleconnections in the geopotential height field during the Northern Hemisphere winter. Mon. Weather Rev. 109, 784–812.

Wallace, J.M., Rasmusson, E.M., Mitchell, T.P., Kousky, V.E., Sarachik, E.S., von Storch, H., 1998. On the structure and evolution of ENSO-related climate variability in the tropical Pacific: lessons from TOGA. J. Geophys. Res. Oceans 103, 14241–14259.

Wang, B., 1995. Interdecadal changes in the El Niño onset in the last four decades. J. Clim. 8, 267–285.

Wang, B., Yang, J., Zhou, T., 2008. Interdecadal changes in the major modes of Asian–Australian Monsoon variability: strengthening relationship with ENSO since the late 1970s. J. Clim. 21, 1771–1789.

Wanner, H., et al., 2001. North Atlantic Oscillation—concepts and studies. Surv. Geophys. 22, 321–382.

Wibig, J., 1999. Precipitation in Europe in relation to circulation patterns at the 500 hPa level. Int. J. Climatol. 19, 253–269.

Widmann, M., Schär, C., 1997. A principal component and long-term trend analysis of daily precipitation in Switzerland. Int. J. Climatol. 17, 1333–1356.

Wilks, D.S., 2006. Statistical Methods in the Atmospheric Sciences. Elsevier (659 pp.).

Wittenberg, A.T., 2009. Are historical records sufficient to constrain ENSO simulations? Geophys. Res. Lett. 36. http://dx.doi.org/10.1029/2009GL038710.

Wunsch, C., 1999. The interpretation of short climate records, with comments on the North Atlantic and Southern Oscillations. BAMS 80, 245–255.

Xie, S.-P., Carton, J.A., 2004. Tropical Atlantic variability: patterns, mechanisms, and impacts. Earth Climate: The Ocean–Atmosphere Interaction, Geophys. Monogr 147. Amer. Geophys. Union, pp. 121–142.

Xie, S.-P., Du, Y., Huang, G., Zheng, X.-T., Tokinaga, H., Hu, K., Liu, Q., 2010. Decadal shift in El Nino influences on Indo-Western Pacific and East Asian climate in the 1970s. J. Clim. 23, 3352–3368.

Zanchettin, D., Timmreck, C., Graf, H.-F., Rubino, A., Lorenz, S., Lohmann, K., Krüger, K., Jungclaus, J.H., 2012. Bi-decadal variability excited in the coupled ocean–atmosphere system by strong tropical volcanic eruptions. Clim. Dyn. 39, 419–444.

Zhou, Z.-Q., Xie, S.-P., Zheng, X.-T., Liu, Q., Wang, H., 2014. Global warming-induced changes in El Niño teleconnections over the North Pacific and North America. J. Clim. 27, 9050–9064.