



North Atlantic climate variability: The role of the North Atlantic Oscillation

James W. Hurrell*, Clara Deser

National Center for Atmospheric Research¹, Climate and Global Dynamics Division, Climate Analysis Section, P.O. Box 3000, Boulder, CO 80307-3000, USA

ARTICLE INFO

Article history:

Received 22 November 2006
Received in revised form 14 May 2007
Accepted 24 November 2008
Available online 20 February 2009

Keywords:

NAO
Climate variability
Physical oceanographic impacts

ABSTRACT

Marine ecosystems are undergoing rapid change at local and global scales. To understand these changes, including the relative roles of natural variability and anthropogenic effects, and to predict the future state of marine ecosystems requires quantitative understanding of the physics, biogeochemistry and ecology of oceanic systems at mechanistic levels. Central to this understanding is the role played by dominant patterns or “modes” of atmospheric and oceanic variability, which orchestrate coherent variations in climate over large regions with profound impacts on ecosystems. We review the spatial structure of extratropical climate variability over the Northern Hemisphere and, specifically, focus on modes of climate variability over the extratropical North Atlantic.

A leading pattern of weather and climate variability over the Northern Hemisphere is the North Atlantic Oscillation (NAO). The NAO refers to a redistribution of atmospheric mass between the Arctic and the subtropical Atlantic, and swings from one phase to another producing large changes in surface air temperature, winds, storminess and precipitation over the Atlantic as well as the adjacent continents. The NAO also affects the ocean through changes in heat content, gyre circulations, mixed layer depth, salinity, high latitude deep water formation and sea ice cover. Thus, indices of the NAO have become widely used to document and understand how this mode of variability alters the structure and functioning of marine ecosystems.

There is no unique way, however, to define the NAO. Several approaches are discussed including both linear (e.g., principal component analysis) and nonlinear (e.g., cluster analysis) techniques. The former, which have been most widely used, assume preferred atmospheric circulation states come in pairs, in which anomalies of opposite polarity have the same spatial structure. In contrast, nonlinear techniques search for recurrent patterns of a specific amplitude and sign. They reveal, for instance, spatial asymmetries between different phases of the NAO that are likely important for ecological studies.

It also follows that there is no universally accepted index to describe the temporal evolution of the NAO. Several of the most common measures are presented and compared. All reveal that there is no preferred time scale of variability for the NAO: large changes occur from one winter to the next and from one decade to the next. There is also a large amount of within-season variability in the patterns of atmospheric circulation of the North Atlantic, so that most winters cannot be characterized solely by a canonical NAO structure. A better understanding of how the NAO responds to external forcing, including sea surface temperature changes in the tropics, stratospheric influences, and increasing greenhouse gas concentrations, is crucial to the current debate on climate variability and change.

© 2009 Elsevier B.V. All rights reserved.

1. Introduction

Ecosystems significantly affect societies and nations by providing essential renewable resources and other benefits, including food, fiber, shelter, energy, biodiversity, clean air and water, recycling of elements, and cultural, spiritual, and aesthetic returns, while human activities, in turn, affect ecosystem processes and dynamics. Ecosystems also affect the climate system by exchanging large amounts of energy, momentum, and greenhouse gases with the atmosphere. Global

climate change is altering the structure and functioning of ecosystems, which in turn affects availability of ecological resources and benefits, changes the magnitude of some feedbacks between ecosystems and the climate system, and could affect economic systems that depend on ecosystems. A grand challenge is to understand and be able to project the potential effects of global climate variability and change on ecosystems, the goods and services ecosystems provide, the drivers and consequences of human responses to ecosystem variability and change, and ecosystem links to the climate system.

Changes in naturally-occurring patterns or “modes” of atmospheric and oceanic variability, such as the El Niño/Southern Oscillation (ENSO) and the North Atlantic Oscillation (NAO), orchestrate large variations in weather and climate over much of the globe on interannual and longer time scales. For instance, a significant

* Corresponding author. Tel.: +1 303 497 1383; fax: +1 303 497 1333.

E-mail address: jhurrell@ucar.edu (J.W. Hurrell).

¹ The National Center for Atmospheric Research is sponsored by the National Science Foundation.

portion of the global warming in recent decades has been attributed to decadal changes in the phase and amplitude of these two dominant patterns of variability. Moreover, it has been argued that the spatial pattern of the response to anthropogenic forcing may project principally onto such modes of natural climate variability (e.g., Corti et al., 1999).

At the same time, there is a growing appreciation that changes in the frequency and amplitude of modes of climate variability profoundly influence a variety of ecological processes and, consequently, temporal and spatial patterns of population and species abundance (e.g., Stenseth et al., 2003; Wang and Schimel, 2003). These changing spatial patterns can have significant societal impacts, for example, by benefiting or harming different nations or groups of resource users, and by disrupting international agreements regarding the division of fishery benefits (Miller and Munro, 2004). Early studies of the influence of climate on ecological systems typically focused on local weather parameters such as temperature, precipitation and snow depth, and this approach is still common in ecology (Stenseth et al., 2005). Indices of large-scale climate modes, in contrast, provide an integrated measure of weather, and therefore can be linked more to the overall physical variability of the system than any individual, local variable. For instance, NAO indices imply information about temperature, storms and precipitation, cloudiness, hydrographic characteristics, mixed-layer depths, and circulation patterns in the ocean that likely explain more of the observed variability of a species than just, for example, water temperature. Moreover, since mode variations produce coherent variations in climate over large regions, they produce impacts on ecosystems at spatial scales that have major effects on society in many ways, including substantial control over Atlantic and Pacific fisheries, terrestrial wildfire and other disturbances, and water availability.

Modal variability thus forms a natural subject on which investigators of climate, ecosystem and climate impact science can collaborate. This collaboration is also required to determine the most relevant effects for society of global change on ecosystems. As part of this special issue, we focus specifically on the extratropical Northern Hemisphere (NH) and begin with a basic description of the spatial structure of extratropical climate and climate variability. This includes a discussion of how the modes of Atlantic climate variability are defined and how they relate to one another. The impacts of the Atlantic variability modes on surface temperature, precipitation, storms, the underlying ocean and sea ice are also briefly described. We conclude by expressing our thoughts on outstanding issues and future challenges, especially related to the mechanisms that most likely govern modal variability.

2. The spatial structure of extratropical climate and climate variability

Climate variability is usually characterized in terms of “anomalies”, where an anomaly is the difference between the instantaneous state of the climate system and the climatology (the mean state computed over many years representative of the era under consideration). Since the spatial structure of climate variability in the extratropics is strongly seasonally dependent (Wallace et al., 1993), it is useful to briefly examine the seasonal evolution of the mean state upon which the climate variations are superimposed.

2.1. The mean state and planetary waves

Large changes in the mean distribution of sea level pressure (SLP) over the NH are evident from winter (December–February) to summer (June–August, Fig. 1). Perhaps most noticeable are those changes over the Asian continent related to the development of the Siberian anticyclone during winter and the monsoon cyclone over Southeast Asia during summer. Over the northern oceans, subtropical

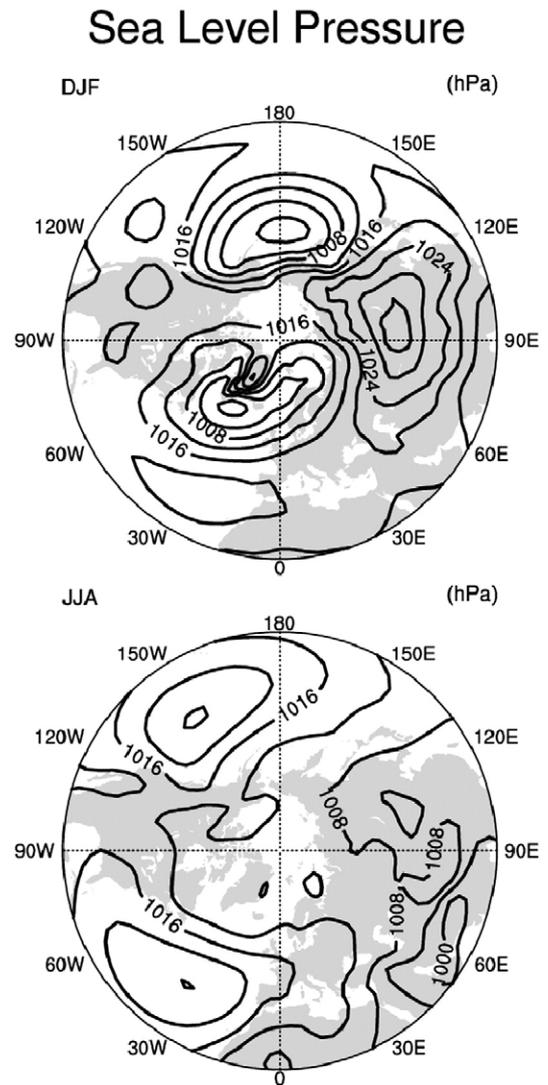


Fig. 1. Mean sea level pressure for (top) boreal winter (December–February) and (bottom) boreal summer (June–August). The data come from the NCEP/NCAR reanalysis project over 1958–2006 (Kalnay et al., 1996) and the contour increment is 4 hPa.

anticyclones dominate during summer, with the Azores high-pressure system covering nearly all of the North Atlantic. These anticyclones weaken and move equatorward by winter, when the high-latitude Aleutian and Icelandic low-pressure centers predominate.

Because air flows counterclockwise around low pressure and clockwise around high pressure in the NH, westerly flow across the middle latitudes of the Atlantic sector occurs throughout the year. The vigor of the flow is related to the north–south pressure gradient, so the surface winds are strongest during winter when they average near 5 m s^{-1} from the eastern United States across the Atlantic onto northern Europe (Fig. 2). These middle latitude westerly winds extend throughout the troposphere and reach their maximum (up to 40 m s^{-1}) at a height of about 12 km. This “jet stream” roughly coincides with the path of storms (atmospheric disturbances operating on time scales of days) traveling between North America and Europe. Over the subtropical Atlantic the prevailing surface northeasterly trade winds are relatively steady but strongest during boreal summer.

In the middle troposphere ($\sim 5\text{--}6 \text{ km}$), the boreal winter map of the geopotential height field reveals a westward tilt with elevation of the high latitude surface cyclones and anticyclones (Fig. 3). Two low-pressure troughs and two high-pressure ridges are evident: the troughs are located over northeastern Canada and just east of Asia, and the ridges are positioned just west of Europe and over western

Vector Winds

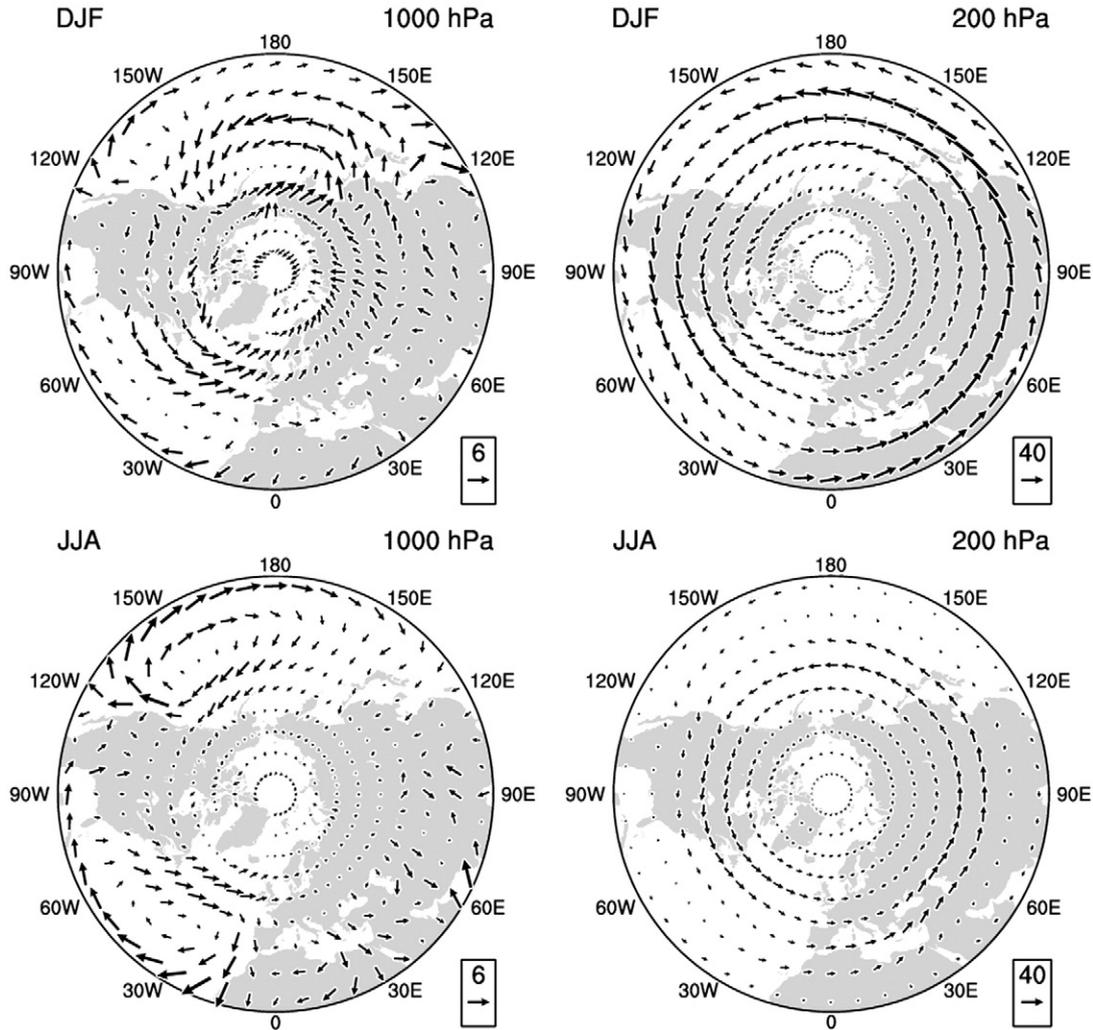


Fig. 2. Mean vector winds for (top) boreal winter (December–February) and (bottom) boreal summer (June–August) for (left) 1000 hPa and (right) 200 hPa over 1958–2006. The scaling vectors in m s^{-1} are indicated in the boxes.

North America. These strong zonal asymmetries reflect the so-called “stationary waves” that are forced primarily by the continent–ocean heating contrasts and the presence of the Rocky and Himalayan mountain ranges. In summer the flow is much weaker and more symmetric, consistent with a much more uniform equator-to-pole distribution of solar radiation.

Although the planetary-scale wave patterns (Fig. 3) are geographically anchored, they do change in time either because the heating patterns in the atmosphere vary or because of internal (chaotic) processes. The amplitude and structure of the variability of the seasonal mean 500 hPa geopotential height field (Fig. 4) is characterized by a strong longitudinal dependence with maximum temporal variance over the northern oceans, especially during boreal winter. The frequency dependence of the winter pattern is subtle: maps of the variability of monthly mean data, or data filtered to retain fluctuations within specific frequency bands (e.g., 60–180 days), also exhibit distinct variance maxima at 500 hPa over the Atlantic and Pacific Oceans, although the longitudinal contrasts become increasingly apparent as longer time scales are examined (Kushnir and Wallace, 1989). In comparison, throughout most of the NH, the standard deviations of boreal summer 500 hPa heights are only about half as large as those of the wintertime means (Fig. 4) (see also Wallace et al., 1993).

2.2. Teleconnections: the PNA and the NAO

A consequence of the transient behavior of the atmospheric planetary-scale waves is that anomalies in climate on seasonal time scales typically occur over large geographic regions. Some regions may be cooler or perhaps drier than average, while at the same time thousands of kilometers away, warmer and wetter conditions prevail. These simultaneous variations in climate, often of opposite sign, over distant parts of the globe are commonly referred to as “teleconnections” in the meteorological literature (Wallace and Gutzler, 1981; Esbensen, 1984; Barnston and Livezey, 1987; Kushnir and Wallace, 1989; Trenberth et al., 1998). Though their precise nature and shape vary to some extent according to the statistical methodology and the data set employed in the analysis, consistent regional characteristics that identify the most conspicuous patterns emerge.

Arguably the most prominent teleconnections over the NH are the NAO and the Pacific–North American (PNA) patterns. Both patterns are of largest amplitude during the boreal winter months, and their mid-tropospheric spatial structure is illustrated most simply through one-point correlation maps (Fig. 5). These maps are constructed by correlating the 500-hPa height time series at a “reference gridpoint” with the corresponding time series at all gridpoints (e.g., Wallace and Gutzler, 1981). The PNA teleconnection pattern has four centers of

500hPa Geopotential Height

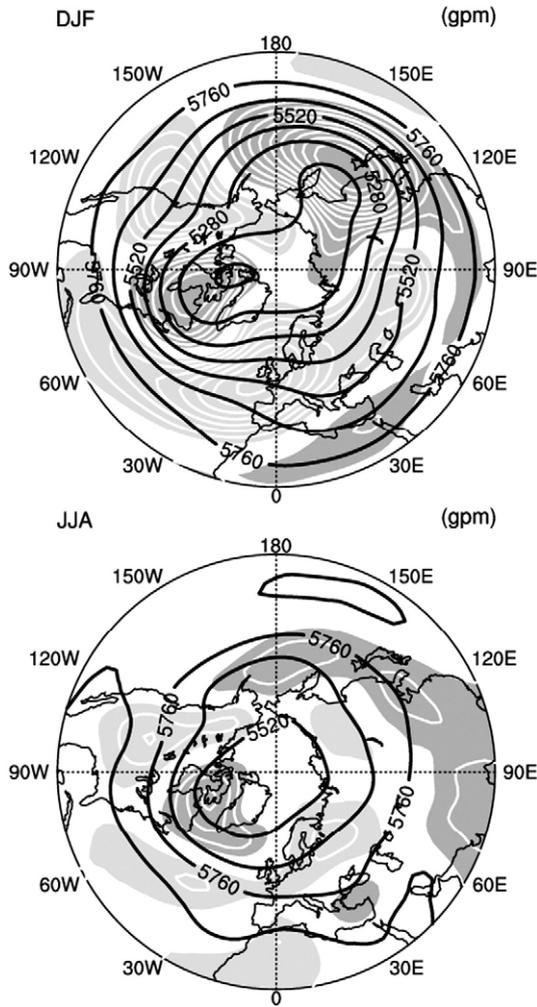


Fig. 3. Mean 500-hPa geopotential height for (top) boreal winter (December–February) and (bottom) boreal summer (June–August), indicated by the thick contours every 120 geopotential meters (gpm), over 1958–2006. The white contours (every 20 gpm, zero contour excluded) indicate departures from the zonal average: negative (positive) departures are indicated by dark (light) shading.

action. Over the North Pacific Ocean, geopotential height fluctuations near the Aleutian Islands vary out-of-phase with those to the south, forming a seesaw pivoted along the mean position of the Pacific subtropical jet stream (Fig. 2). Over North America, variations in geopotential height over western Canada and the northwestern U.S. are negatively correlated with those over the southeastern U.S., but are positively correlated with the subtropical Pacific center. The significance of the locations and the respective phases of the four centers of the PNA is their relation to the mean atmospheric circulation (Fig. 3). Variations in the PNA pattern represent variations in the waveness of the atmospheric flow in the western half-hemisphere and thus changes in the north–south migration of the large-scale Pacific and North American air masses and their associated weather.

On interannual time scales, atmospheric circulation anomalies over the North Pacific, including the PNA, are linked to changes in tropical Pacific sea surface temperatures associated with the El Niño/Southern Oscillation (ENSO) phenomenon. This association reflects mainly the dynamical teleconnection to higher latitudes forced by deep convection in the tropics (see Trenberth et al., 1998 for a review). The PNA pattern is sometimes viewed, then, as the extratropical arm

of ENSO. Significant variability of the PNA occurs even in the absence of ENSO, however, indicating that the PNA is an “internal” mode of atmospheric variability.

Similarly, the NAO does not owe its existence to coupled ocean–atmosphere–land interactions (Thompson et al., 2003; Czaja et al., 2003), as is evident from climate model experiments that do not include SST, sea ice or land surface variability (Hurrell et al., 2003). In contrast to the wave-like appearance of the PNA, the NAO is primarily a north–south dipole characterized by simultaneous out-of-phase height anomalies between temperate and high latitudes over the Atlantic sector (Fig. 5). Both the NAO and PNA are also reflected in the spatial patterns of the two leading empirically-determined orthogonal functions (EOFs) of NH boreal winter 500 hPa height (not shown). The NAO also dominates the leading EOF of the NH SLP field (Kutzbach, 1970; Rogers, 1981; Trenberth and Paolino, 1981; Thompson et al., 2003). Analyzing SLP allows for the longer-term behavior of the NAO to be evaluated, as a long series of SLP charts over the NH begin in 1899 (Trenberth and Paolino, 1980), in contrast to 500 hPa height charts that are confined to after 1947. Moreover, even longer instrumental records of SLP variations are available, especially from European stations (Jones et al., 2003). Thus, in the following, we

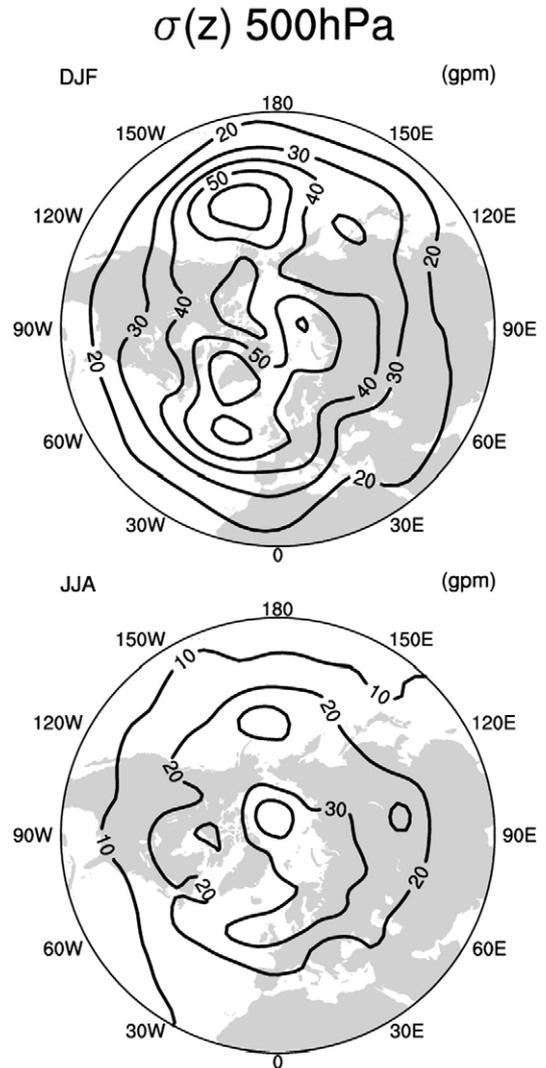


Fig. 4. Interannual variability of 500-hPa geopotential height for (top) boreal winter (December–February) and (bottom) boreal summer (June–August) over 1958–2006. The contour increment is 10 gpm.

One-Point Correlation

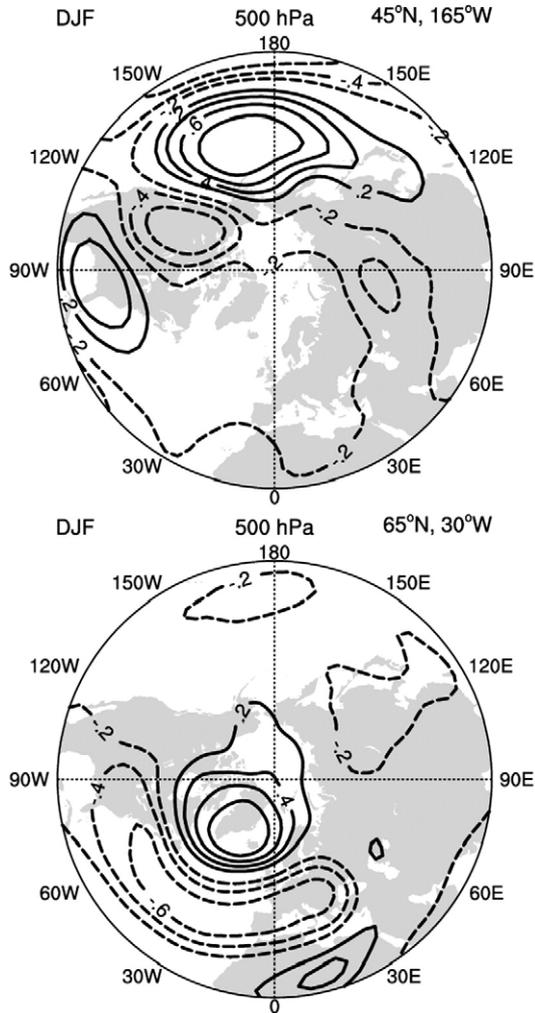


Fig. 5. One-point correlation maps of 500 hPa geopotential heights for boreal winter (December–February) over 1958–2006. In the top panel, the reference point is 45°N, 165°W, corresponding to the primary center of action of the PNA pattern. In the lower panel, the NAO pattern is illustrated based on a reference point of 65°N, 30°W. Negative correlation coefficients are dashed, the contour increment is 0.2, and the zero contour has been excluded.

examine the spatial structure and time evolution of the NAO in more detail from SLP records.

3. The spatial signature of the NAO

There is no single way to “define” the NAO. One approach is through conceptually simple one-point correlation maps (e.g., Fig. 5), identifying the NAO by regions of maximum negative correlation over the North Atlantic (Wallace and Gutzler, 1981; Kushnir and Wallace, 1989; Portis et al., 2001). Another technique is EOF (or principal component) analysis. In this approach, the NAO is identified from the eigenvectors of the cross-covariance (or cross-correlation) matrix, computed from the time variations of the gridpoint values of SLP or some other climate variable. The eigenvectors, each constrained to be spatially and temporally orthogonal to the others, are then scaled according to the amount of total data variance they explain. This linear approach assumes preferred atmospheric circulation states come in pairs, in which anomalies of opposite polarity have the same spatial structure. In contrast, climate anomalies can also be identified by cluster analysis techniques, which search for recurrent patterns of a

specific amplitude and sign. Clustering algorithms identify weather or climate “regimes”, which correspond to peaks in the probability density function of the climate phase space (Lorenz, 1963). Interest in this nonlinear interpretation of atmospheric variability has been growing, and recently has found applications within the climate framework (e.g., Palmer, 1999; Corti et al., 1999; Cassou and Terray, 2001a,b; Cassou et al., 2004a). In the following, we compare the spatial patterns of the NAO as estimated from both traditional EOF and clustering techniques.

3.1. EOF analysis of North Atlantic SLP

The leading eigenvectors of the cross-covariance matrix calculated from seasonal (3-month average) SLP anomalies in the North Atlantic sector (20°–70°N; 90°W–40°E) are illustrated in Fig. 6. The patterns are very similar if based on the cross-correlation matrix (not shown).

The largest amplitude anomalies in SLP occur during the boreal winter months; however, throughout the year the leading pattern of variability is characterized by a surface pressure dipole, and thus may be viewed as the NAO, although the spatial pattern is not stationary (Barnston and Livezey, 1987; Hurrell and van Loon, 1997; Portis et al., 2001). Since the eigenvectors are, by definition, structured to explain maximum variance, it is expected that the “center of actions” of the leading EOFs will coincide with the regions of strongest variability, and the movement of those regions through the annual cycle is reflected in Fig. 6.

The NAO is the only teleconnection pattern evident throughout the year in the NH (Barnston and Livezey, 1987). During the winter season (December–February), it accounts for more than one-third of the total variance in SLP over the North Atlantic, and appears with a slight northwest-to-southeast orientation. In the so-called “positive phase” (depicted), higher-than-normal surface pressures south of 55°N combine with a broad region of anomalously low pressure throughout the Arctic to enhance the climatological meridional pressure gradient (Fig. 1). The largest amplitude anomalies occur in the vicinity of Iceland and across the Iberian Peninsula. The positive phase of the NAO is associated with stronger-than-average surface westerlies across the middle latitudes of the Atlantic onto Europe, with anomalous southerly flow over the eastern U.S. and anomalous northerly flow across the Canadian Arctic and the Mediterranean (Fig. 7).

The NAO is well separated (and thus less likely to be affected by statistical sampling errors) in all seasons from the second eigenvector, according to the criterion of North et al. (1982). By boreal spring (March–May), the NAO appears as a north–south dipole with a southern center of action near the Azores. Both the spatial extent and the amplitude of the SLP anomalies are smaller than during winter, but not by much, and the leading EOF explains 30% of the SLP variance. The amplitude, spatial extent, and the percentage of total SLP variability explained by the NAO reach minimums during the summer (June–August) season, when the centers of action are substantially north and east relative to winter. By fall (September–November), the NAO takes on more of a southwest-to-northeast orientation, with SLP anomalies in the northern center of action comparable in amplitude to those during spring.

That the spatial pattern of the NAO remains largely similar throughout the year does not imply that it also tends to persist in the same phase for long. To the contrary, it is highly variable, tending to change its phase from one month to another, and its longer-term time-average behavior reflects the combined effect of residence time in any given phase and its amplitude therein.

Most studies of the NAO focus on the NH winter months, when the atmosphere is most active dynamically and perturbations grow to their largest amplitudes. As a result, the influence of the NAO on surface temperature and precipitation is likely to be greatest at this time of year, and we therefore focus on the winter variations.

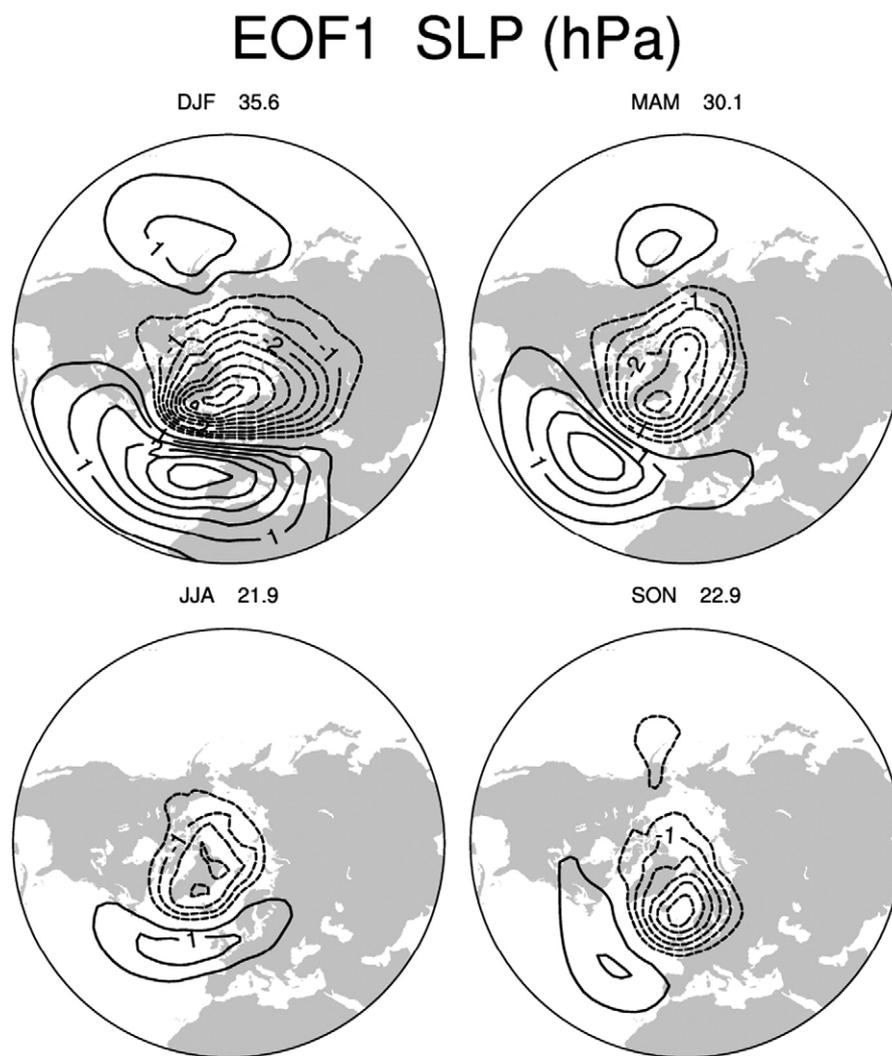


Fig. 6. Leading empirical orthogonal functions (EOF 1) of the seasonal mean sea level pressure anomalies in the North Atlantic sector (20° – 70° N, 90° W– 40° E), and the percentage of the total variance they explain. The patterns are displayed in terms of amplitude (hPa), obtained by regressing the hemispheric sea level pressure anomalies upon the leading principal component time series. The contour increment is 0.5 hPa, and the zero contour has been excluded. The data cover 1899–2006 (see Trenberth and Paolino, 1980).

However, coherent fluctuations of surface pressure, temperature, cloudiness and precipitation occur throughout the year over the North Atlantic, and decadal and longer-term variability is not confined to winter (Cassou et al., 2004b). Moreover, the vigorous wintertime NAO can interact with the slower components of the climate system (the ocean, in particular) to leave persistent surface anomalies into the ensuing parts of the year that may significantly influence the evolution of the climate and marine ecosystems (Czaja et al., 2003; Rodwell, 2003).

3.2. EOF analysis of NH SLP

A well-known shortcoming of EOF analysis is that eigenvectors are mathematical constructs, constrained by their mutual orthogonality and the maximization of variance over the entire analysis domain. There is no guarantee, therefore, that they represent physical/dynamical modes of the climate system. An EOF analysis, for instance, will not clearly reveal two patterns that are linearly superposed if those patterns are not orthogonal. Moreover, the loading values of EOFs do not reflect the local behavior of the data: values of the same sign at two different spatial points in an EOF do not imply that those two points are significantly correlated. This means that the pattern structure of any particular EOF must be interpreted with care (e.g., Dommenges and Latif, 2002). These issues have been at the center of a

debate (e.g., Deser, 2000) over whether or not the NAO is a regional expression of a larger-scale (hemispheric) mode of variability known as the NH Annular Mode (NAM; Thompson et al., 2003) or, previously, the Arctic Oscillation.

The NAM is defined as the first EOF of NH (20° – 90° N) winter SLP data (Fig. 8, upper panel, based on the cross-covariance matrix). It explains 23% of the extended winter-mean (December–March) variance, and it is clearly dominated by the NAO structure in the Atlantic sector. Although there are some subtle differences from the regional pattern (Fig. 8, lower panel) over the Atlantic and Arctic, the main difference is larger amplitude anomalies over the North Pacific of the same sign as those over the Atlantic. This feature gives the NAM an almost annular (or zonally-symmetric) structure that reflects a more hemispheric-scale meridional seesaw in SLP between polar and middle latitudes. Though first identified by Lorenz (1951) in zonally-averaged data and by Kutzbach (1970), Wallace and Gutzler (1981), and Trenberth and Paolino (1981) in gridded data, Thompson and Wallace (1998, 2000) have argued that the NAM is a fundamental structure of NH climate variability, and that the “regional” NAO reflects the modification of the annular mode by zonally-asymmetric forcings, such as topography and land–ocean temperature contrasts. It would then follow that the annular mode perspective is critical in order to understand the processes that give rise to NAM (or NAO) variations. For instance, the leading wintertime pattern of variability

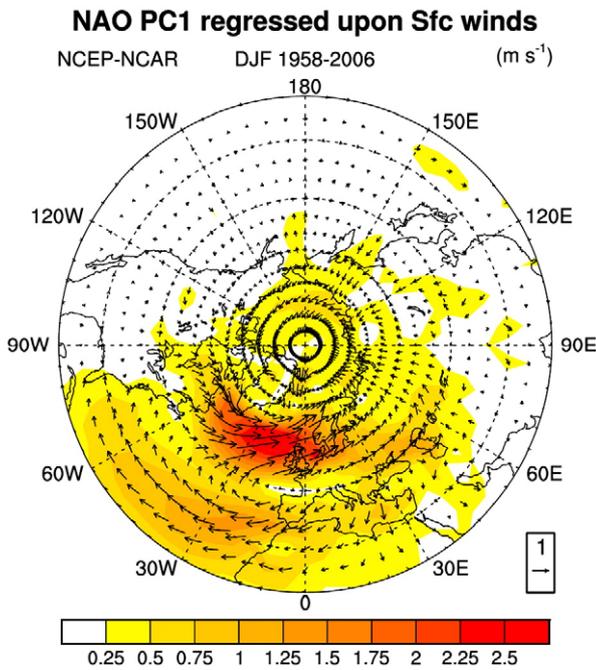


Fig. 7. The amplitude of boreal winter (December–February) 1000 hPa scalar wind speed (color) and vector wind (arrows) associated with a one standard deviation change of an NAO index, defined as the principal component time series of the leading empirical orthogonal function of Atlantic-sector sea level pressure (as in Fig. 6). The plot is constructed from winter data from the NCEP/NCAR reanalyses over 1958–2006. The color scale is in units of m s^{-1} , and the scaling vector (lower right) is 1 m s^{-1} .

in the lower stratosphere is clearly annular (not shown), but the SLP anomaly pattern that is associated with it is confined almost entirely to the Arctic and Atlantic sectors and coincides with the spatial structure of the NAO (e.g., Deser, 2000). Thompson et al. (2003) present a thorough overview of the dynamics governing annular mode behavior, including a discussion of the mechanisms by which annular variability in the stratosphere might drive NAO-like variations in surface climate.

3.3. Cluster analysis of North Atlantic SLP

The dynamical signature of interannual variability in the North Atlantic domain can also be examined through non-linear approaches, such as cluster analysis or non-linear principal component analysis (Monahan et al., 2000, 2001). Here we apply the former to 57 years of daily SLP data from December–March (DJFM) using the procedures of Cassou and Terray (2001a,b) and Cassou et al. (2004a), which are based on the clustering algorithm of Michelangeli et al. (1995). The solutions are robust among different algorithms and SLP data sets (not shown). Briefly, cluster analysis is a multivariate statistical technique that groups together the daily SLP maps into a small number of representative states (or regimes) according to an objective criterion of similarity. Note that by construction, the percentage of occurrence of the identified clusters sums to 100. Further information on cluster analysis may be found in Cassou et al. (2004a).

The clustering algorithm applied over the Atlantic domain (20° – 70°N ; 90°W – 40°E) identifies four winter climate regimes in SLP (Fig. 9; see also Cassou et al., 2004a). Two of them correspond to the negative and positive phases of the NAO, while the third and fourth regimes display strong anticyclonic ridges over Scandinavia (the “Blocking” regime) and off western Europe (the “Atlantic Ridge” regime). The latter bears some resemblance to another prominent atmospheric teleconnection: the East Atlantic pattern (e.g., Wallace and Gutzler, 1981; Barnston and Livezey, 1987). All four regimes occur with about the same frequency (20–30% of all winter days), although these numbers are sensitive to the period of analysis, reflecting that

the dominance of certain regimes over others varies over time (see Section 4).

In contrast to the typical NAO pattern identified through linear approaches (e.g., Figs. 5 and 6), some interesting spatial asymmetries are evident in Fig. 9. Most striking is the difference in the position of the pressure anomalies between the two NAO regimes: in particular, the eastward shift (by $\sim 30^{\circ}$ longitude) and northeastward extension of the subpolar SLP anomalies in the positive relative to the negative regime (see also Cassou et al., 2004a). These spatial asymmetries are not dependent on the analysis period: they are evident in subperiods of the SLP data set. Similar results, indicating a non-linearity in NAO variability, are found when the PC time series of the leading EOF of Atlantic SLP is used to define and average together positive and negative index winters (not shown). The robustness of the eastward displacement of the NAO in positive regime months has interesting implications for conclusions drawn from some climate model studies on how increasing greenhouse gas (GHG) concentrations might affect the spatial structure of the NAO (Gillett, 2003). Ulbrich and Christoph (1999), for instance, concluded that future enhanced GHG forcing

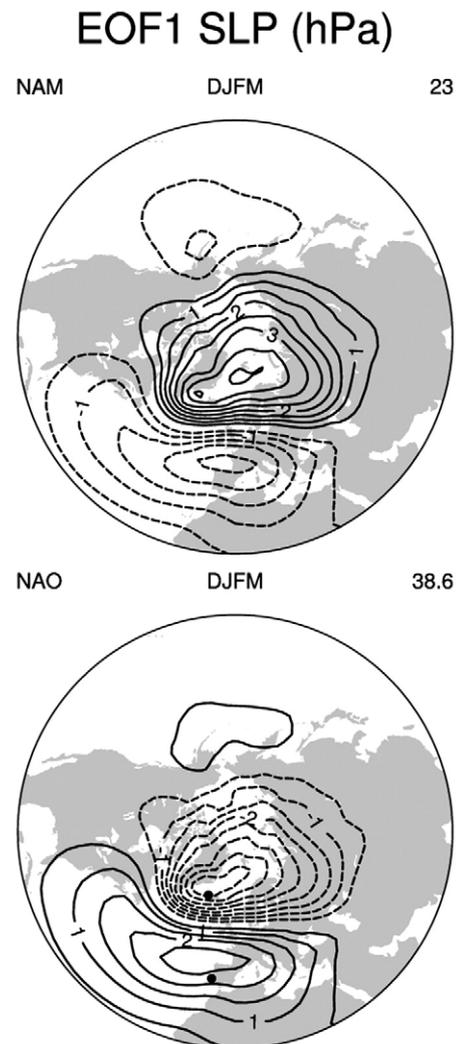


Fig. 8. Leading empirical orthogonal function (EOF 1) of the winter (December–March) mean sea level pressure anomalies over (top) the Northern Hemisphere (20° – 90°N) and (bottom) the North Atlantic sector (20° – 70°N , 90°W – 40°E), and the percentage of the total variance they explain. The patterns are displayed in terms of amplitude (hPa), obtained by regressing the hemispheric sea level pressure anomalies upon the leading principal component time series. The contour increment is 0.5 hPa, and the zero contour has been excluded. The data cover 1899–2006. The dots in the bottom panel represent the locations of Lisbon, Portugal and Stykkisholmur, Iceland used in the station based NAO index of Hurrell (1995; see Fig. 10).

4 SLP Weather Regimes for NCEP (DJFM 1950-2006)

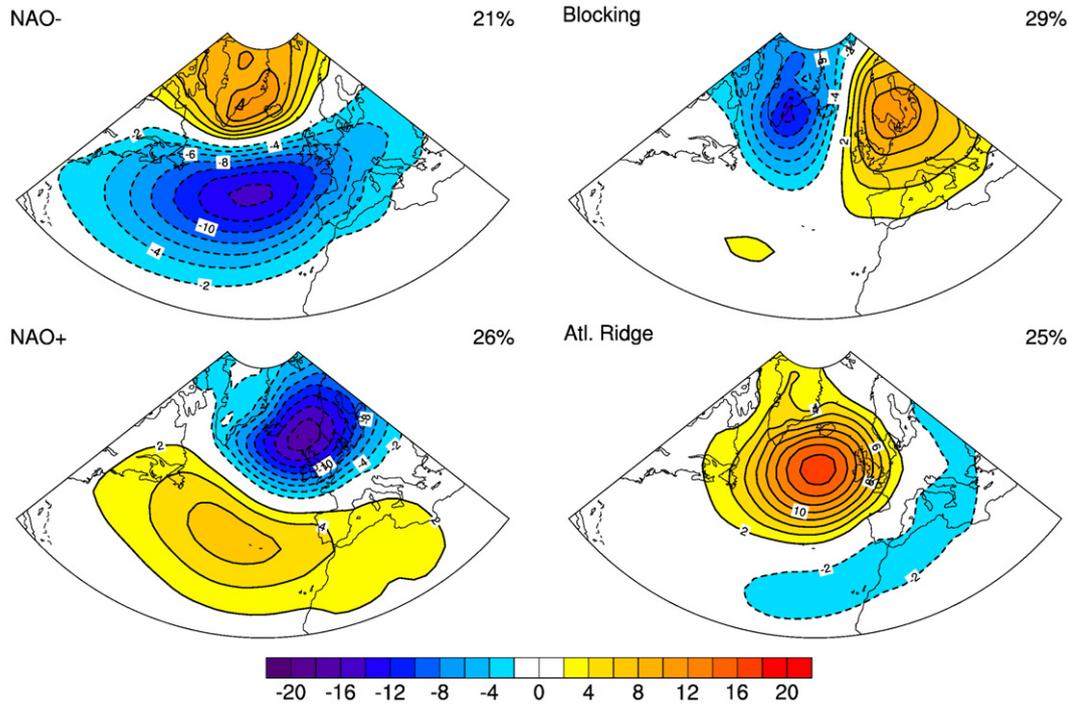


Fig. 9. Boreal winter (December–March) climate regimes in sea level pressure (hPa) over the North Atlantic domain (20°–70°N, 90°W–40°E) using daily data over 1950–2006. The percentage at the top right of each panel expresses the frequency of occurrence of a cluster out of all winter days since 1950. The contour interval is 2 hPa.

might result in an eastward displacement of the NAO centers of action. The results from the regime analysis, however, suggest that longitudinal shifts could arise from the preferential excitation of positive NAO regimes, which are intrinsically displaced eastward, rather than a static shift of the Atlantic pressure centers. Hilmer and Jung (2000) documented an eastward shift of the centers of interannual NAO variability over the period 1978–1997 relative to 1958–1977, and they postulated that such a change could have arisen from a change in the occupation statistics of fixed modes (see also Lu and Greatbatch, 2002). As we show below, this seems to be the case.

4. Temporal variability of the NAO

Since there is no unique way to define the spatial structure of the NAO, it follows that there is no universally accepted index to describe the temporal evolution of the phenomenon. Most modern NAO indices are derived either from the simple difference in surface pressure anomalies between various northern and southern locations, or from the PC time series of the leading (usually regional) EOF of SLP. Many examples of the former exist, usually based on instrumental records from individual stations near the NAO centers of action (e.g., Rogers, 1984; Hurrell, 1995; Jones et al., 1997; Slonosky and Yiou, 2001), but sometimes from gridded SLP analyses (e.g., Portis et al., 2001; Luterbacher et al., 2002). Jones et al. (2003) discuss and compare various station-based indices in detail.

A disadvantage of station-based indices is that they are fixed in space. Given the movement of the NAO centers of action through the annual cycle (Fig. 6), such indices can only adequately capture NAO variability for parts of the year (Hurrell and van Loon, 1997; Portis et al., 2001; Jones et al., 2003). Moreover, individual station pressures are significantly affected by small-scale and transient meteorological phenomena not related to the NAO and, thus, contain noise (see Trenberth, 1984). Hurrell and van Loon (1997) showed, for instance, that the signal-to-noise ratio of commonly-used winter NAO station-based indices is near 2.5, but by summer it falls to near unity.

An advantage of the PC time series approach is that such indices are more optimal representations of the full NAO spatial pattern; yet, as they are based on gridded SLP data, they can only be computed for short (relative to some station records) periods of time, depending on the data source. Below we compare a station-based index to the PC time series of the leading EOF (PC1) of both Atlantic-sector and NH SLP. The latter is the NAM index of Thompson and Wallace (1998). We also present the time history of occurrence of the NAO regimes identified in Fig. 9. All comparisons are for the winter (December–March) season.

The winter-mean NAO station index of Hurrell (1995) is shown in Fig. 10 (upper panel). Positive values of the index indicate stronger-than-average westerlies over the middle latitudes. The station-based index for the winter season agrees well with PC1 of Atlantic-sector SLP: the correlation coefficient between the two is 0.92 over the common period 1899–2005, indicating that the station-based index adequately represents the time variability of the winter-mean NAO spatial pattern. Moreover, it correlates with the NAM index (lower panel) at 0.85, while the correlation of the two PC1 time series is 0.95. These results again emphasize that the NAO and NAM reflect essentially the same mode of tropospheric variability. When intra-seasonal anomalies are considered by stringing together the individual winter months, the correlation coefficient between the two PC1 time series is reduced slightly to 0.89.

An important conclusion from Fig. 10 is that there is little evidence for the NAO to vary on any preferred time scale. Large changes can occur from one winter to the next, as well as from one decade to the next. The power spectra of the indices in Fig. 10 are only slightly “red”, with power increasing with period (not shown). Feldstein (2000) examined the spectral characteristics of the NAO using daily data and concluded that its temporal evolution is generally consistent with a stochastic (Markov, or first-order autoregressive) process with a fundamental time scale of about 10 days. This then means that observed interannual and longer time scale NAO fluctuations (Fig. 10) could primarily be a statistical remnant of the energetic weekly variability. This “climate noise paradigm” fails, however, to explain the enhanced interannual NAO

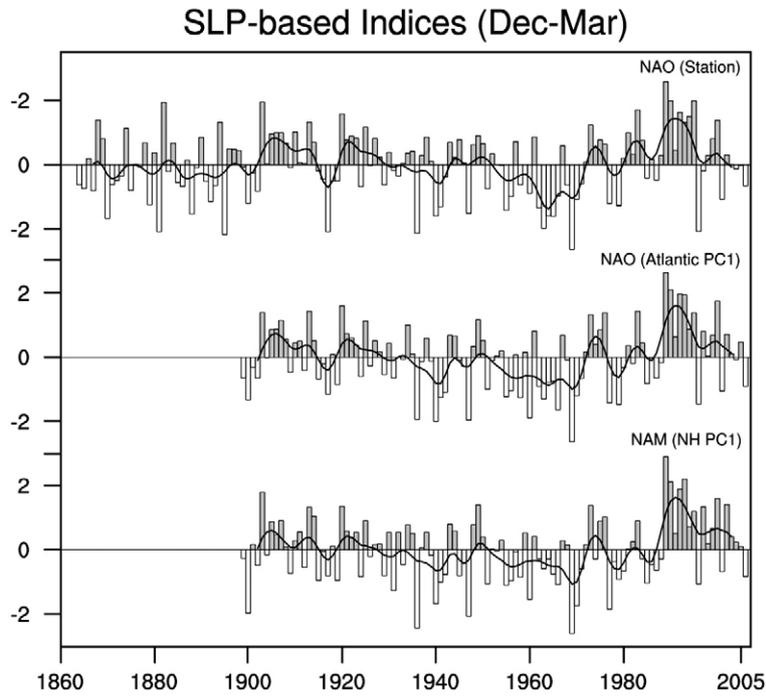


Fig. 10. Normalized indices of the mean winter (December–March) NAO constructed from sea level pressure data. In the top panel, the index is based on the difference of normalized sea level pressure between Lisbon, Portugal and Stykkisholmur/Reykjavik, Iceland. The average winter sea level pressure data at each station were normalized by division of each seasonal pressure by the long-term mean (1864–1983) standard deviation. In the middle and lower panels, the index is the principal component time series of the leading empirical orthogonal function (EOF) of Atlantic-sector and northern hemisphere sea level pressure, respectively. The heavy solid lines represent the indices smoothed to remove fluctuations with periods less than 4 years. The indicated year corresponds to the January of the winter season (e.g. 1990 is the winter of 1989/1990).

variability observed over the last half of the 20th century (Feldstein, 2000, 2002), when a role for forcing by other climate system components is likely. Indeed, numerous studies have argued that variations in heat exchange between the atmosphere and ocean, sea-ice

and/or land systems could modulate NAO variability on seasonal to multi-decadal time scales (Hurrell et al., 2003).

Another index, the time history of the occurrence of NAO, Blocking and Atlantic Ridge regimes (Fig. 8), offers a different perspective (Fig. 11).

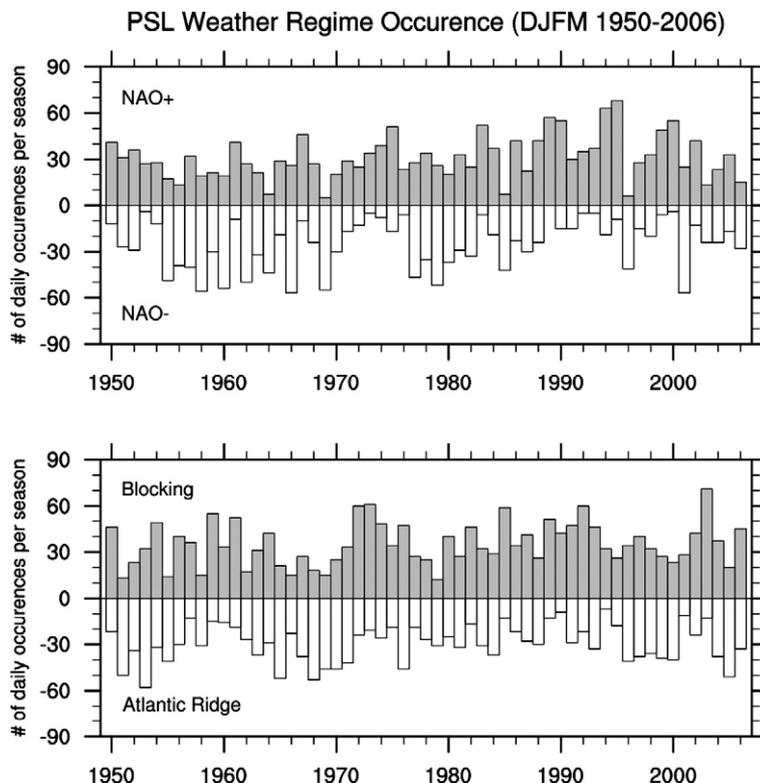


Fig. 11. The time history of occurrence of the NAO, Atlantic Ridge and Blocking regimes (see Fig. 9) over 1950–2006. The vertical bars give the number of days in each winter (December–March) season that the given regime is present. The indicated year corresponds to the January of the winter season (e.g. 1990 is the winter of 1989/1990).

Plotted is frequency of occurrence of each regime in units of the number days a regime is present within a given winter (December–March) season. For the two NAO regimes, as for the more conventional indices, strong interannual variability is evident, and there are periods when one NAO regime occurs almost to the exclusion of the other. For instance, very few positive NAO regime occurrences are found during the 1960s, while very few negative regime occurrences were observed during the 1990s, consistent with the upward trend in traditional NAO indices over this period (Fig. 10).

The regime analysis also illustrates two other important points. First, there is a large amount of within-season variance in the atmospheric circulation of the North Atlantic. Most winters are not dominated by any particular regime; rather, the atmospheric circulation anomalies in one month might resemble the positive index phase of the NAO, while in another month they resemble the negative index phase or some other pattern altogether. Since 2001, for instance, more winter days over the North Atlantic have been characterized by circulation anomalies that project onto the Atlantic Ridge or Blocking patterns than either phase of the NAO (Fig. 11). Moreover, roughly the same numbers of negative and positive NAO index days have occurred over this period, consistent with the small, winter-mean values of conventional NAO indices (Fig. 10). Thus, the second point is that 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 characterized solely by the canonical NAO pattern in Fig. 6.

5. Impacts of the NAO

The NAO exerts a dominant influence on wintertime temperatures across much of the NH. Surface air temperature and SST across wide regions of the North Atlantic Ocean, North America, the Arctic, Eurasia and the Mediterranean are significantly correlated with NAO variability. These changes, along with related changes in storminess and precipitation, ocean heat content, ocean currents and their related heat transport, and sea ice cover have significant impacts on a wide range of human activities as well as on marine, freshwater and terrestrial ecosystems. In the following, we present a very brief overview of these impacts, with an emphasis on those of relevance to marine ecosystems. More detailed discussions are contained elsewhere, including impacts on terrestrial (e.g., [Mysterud et al., 2003](#)) and fresh water ([Straile et al., 2003](#)) ecosystems.

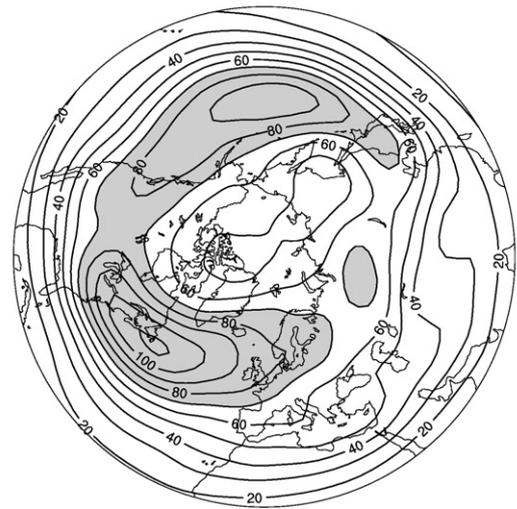
It is important to keep in mind that a particular impact may depend crucially on the local or regional details of the atmospheric circulation pattern that is forcing it, and that these details may not always be adequately reflected in a simple index such as the NAO. For instance, the export of sea ice through Fram Strait is sensitive to the cross-strait SLP gradient and associated northerly wind component. This cross-strait SLP gradient correlates strongly with the NAO in some periods and weakly in others due to slight shifts in the spatial pattern of the NAO through time. Thus, the association between ice export through Fram Strait and the NAO is non-stationary ([Hilmer and Jung, 2000](#)).

5.1. Storms and precipitation

Changes in the mean circulation patterns over the North Atlantic associated with the NAO are accompanied by changes in the intensity and number of storms, their paths, and their weather. During winter, a well-defined storm track connects the North Pacific and North Atlantic basins, with maximum storm activity over the oceans (Fig. 12). The details of changes in storminess differ depending on the analysis method and whether one focuses on surface or upper-air features. Generally, however, positive NAO index winters are associated with a northeastward shift in the Atlantic storm activity with enhanced activity from Newfoundland into northern Europe and a modest

Mean Storm Track (DJFM) 1958–1998

$(Z'^2)^{1/2}$ 300 hPa (gpm)



Regression onto NAO Index (gpm)



Fig. 12. In the top panel, mean storm tracks for 1958–1998 winters (December–March) as revealed by the 300 hPa root mean square transient geopotential height (gpm) bandpassed to include 2–8 day period fluctuations. Values greater than 70 gpm are shaded and the contour increment is 10 gpm. In the lower panel, anomalies are expressed in terms of amplitude (gpm) regressed onto the NAO index (defined as in the middle panel of Fig. 10). The contour increment is 2 gpm, and anomalies greater than 4 gpm in magnitude are shaded. The data come from the NCEP/NCAR reanalyses.

decrease in activity to the south ([Rogers, 1990, 1997](#); [Hurrell and van Loon, 1997](#); [Serreze et al., 1997](#); [Alexandersson et al., 1998](#)). Positive NAO index winters are also typified by more intense and frequent storms in the vicinity of Iceland and the Norwegian Sea ([Serreze et al., 1997](#); [Deser et al., 2000](#)).

The ocean exhibits a marked response to long lasting shifts in the storm climate. For instance, the very persistent and positive NAO index winters of the 1990s were associated with increased wave heights over the northeast Atlantic and decreased wave heights south of 40°N ([Bacon and Carter, 1993](#); [Kushnir et al., 1997](#); [Carter, 1999](#)). Such changes have consequences for the regional ecology, as well as for the operation and safety of shipping, offshore industries such as oil and gas exploration, and coastal development.

Changes in the mean flow and storminess associated with swings in the NAO index are also reflected in pronounced changes in the transport and convergence of atmospheric moisture and, thus, the

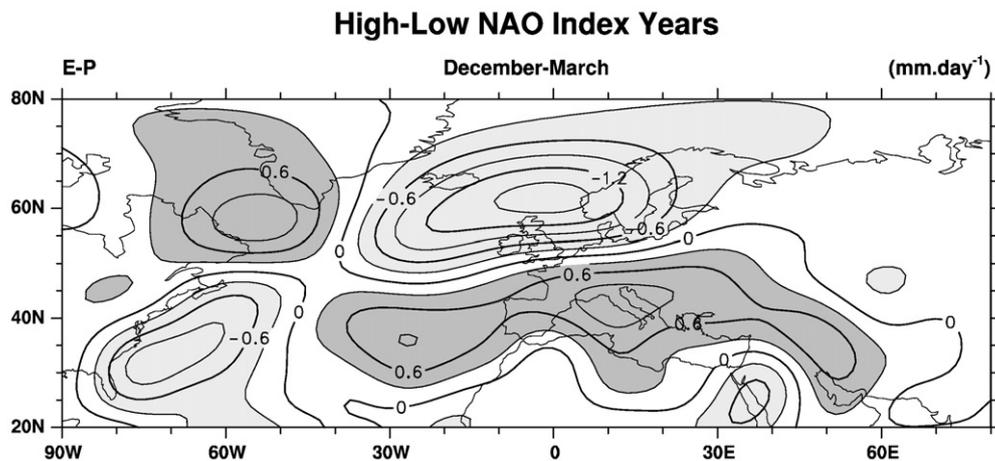


Fig. 13. Difference in mean winter (December–March) evaporation (E) minus precipitation (P) between years when the NAO index exceeds one standard deviation and years when it is less than -1 standard deviation. The NAO index is defined as in the middle panel of Fig. 10, and nine winters enter into both the high index and the low index composites. The E–P field is obtained as a residual of the atmospheric moisture budget (see Hurrell, 1995). The calculation was based on the NCEP/NCAR reanalyses over 1958–2001, and truncated to 21 wavenumbers. The contour increment is 0.3 mm day^{-1} , differences greater than 0.3 mm day^{-1} (E exceeds P) are indicated by dark shading, and differences less than -0.3 mm day^{-1} (P exceeds E) are indicated by light shading.

distribution of evaporation and precipitation (Hurrell, 1995; Dickson et al., 2000). Evaporation exceeds precipitation over much of Greenland and the Canadian Arctic during high NAO index winters (Fig. 13), where changes between high and low NAO index states are on the order of 1 mm day^{-1} . Drier conditions of the same magnitude also occur over much of central and southern Europe, the Mediterranean and parts of the Middle East, whereas more precipitation than normal falls from Iceland through Scandinavia (Hurrell, 1995; Dai et al., 1997; Dickson et al., 2000; Visbeck et al., 2003).

5.2. SST

It has long been recognized that fluctuations in SST and the strength of the NAO are related. The leading pattern of SST variability during boreal winter (not shown) consists of a tri-polar structure marked, in one phase, by a cold anomaly in the subpolar North Atlantic, a warm anomaly in the middle latitudes centered off Cape Hatteras, and a cold subtropical anomaly between the equator and 30°N . This structure suggests the SST anomalies are driven by changes in the air–sea heat exchanges and surface wind induced Ekman currents associated with NAO variations (Marshall et al., 2001a; Visbeck et al., 2003). The relationship is indeed strongest when the NAO index leads an index of the SST variability by several weeks, which highlights the well-known result that large-scale SST over the extratropical oceans responds to atmospheric forcing on monthly and seasonal time scales (e.g., Cayan, 1992a,b; Battisti et al., 1995; Delworth, 1996; Deser and Timlin, 1997). Compositing North Atlantic SST on high and low NAO index winters clearly illustrates the aforementioned tri-pole pattern of SST change (Fig. 14).

Over longer periods, persistent SST anomalies also appear to be related to persistent anomalous patterns of SLP, including those associated with the NAO, but the mechanisms whereby the atmosphere forces SST anomalies on decadal and longer time scales are different from those on interannual time scales (see Bjerknes, 1964 for a pioneering discussion of this issue). On decadal and longer time scales, the ocean adjusts dynamically to the overlying changes in wind stress curl, both locally *via* Ekman pumping and non-locally *via* changes in the gyre-scale circulation (Marshall et al., 2001b; Visbeck et al., 2003). This dynamical adjustment alters the horizontal and vertical oceanic heat transports, which in turn impact SST. It is quite likely, for instance, that sustained NAO forcing results in a basin-wide SST response in which the northern and subtropical parts of the tri-polar pattern merge (Visbeck et al., 2003). There is also evidence for a northward shift in the position of the Gulf Stream during the positive

phase of the NAO, consistent with Sverdrup adjustment of the ocean gyre circulation.

5.3. Subsurface ocean changes

Subsurface ocean observations more clearly depict long-term climate variability, because the effect of the annual cycle and month-to-month variability in the atmospheric circulation decays rapidly with depth. These measurements are much more limited than surface observations, but over the North Atlantic they too indicate fluctuations that are coherent with the low frequency winter NAO index to depths of 400 m (Curry and McCartney, 2001).

Oceanic mixed-layer depth (MLD) is an important physical factor influencing marine biological productivity and ecosystem dynamics. The depth of the mixed-layer determines both the amount of sunlight and the concentration of nutrients available to phytoplankton for photosynthesis. MLD is influenced by atmospheric and oceanic conditions through wind-induced vertical mixing, heat exchange, and upwelling. The response of the winter MLD to the NAO, estimated by compositing winter months (December–March) over 1955–2003 using data from White (1995), consists of a region of negative anomalies (shallower-than-normal mixed layers) extending across much of the middle North Atlantic from the southeastern U.S. to Spain

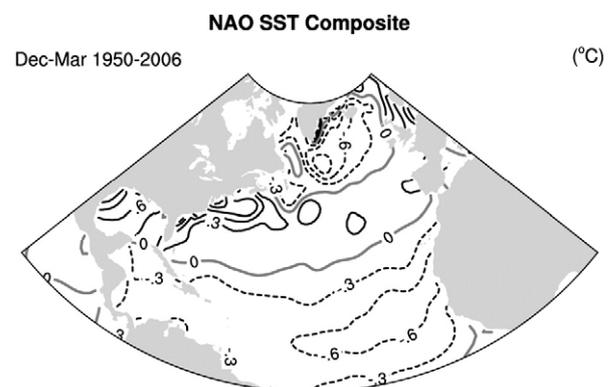


Fig. 14. Difference in mean winter (December–March) sea surface temperature between years when the NAO index exceeds one standard deviation over 1950–2006. The NAO index is defined as in the middle panel of Fig. 10. The contour increment is $0.3 \text{ }^\circ\text{C}$ and positive (negative) differences are given by the solid (dashed) contours.

and a region of weaker positive anomalies to the south (Fig. 15; note that there are insufficient data to determine NAO-related MLD variations in the subpolar North Atlantic.) This pattern of MLD anomalies is similar to the pattern of SST anomalies associated with the NAO (Fig. 14), with negative MLD anomalies corresponding to positive SST anomalies and *vice versa*.

The SST tri-pole pattern (Fig. 14) has been shown to recur from one winter to the next with little persistence during the intervening summer (Watanabe and Kimoto, 2000; Timlin et al., 2002; Deser et al., 2003; de Coëtlogon and Frankignoul, 2003). The mechanism for this winter-to-winter memory of the SST anomaly tri-pole is due to the seasonal cycle of MLD through the so-called “reemergence mechanism” (Alexander and Deser, 1995). Briefly, the winter NAO creates a tri-pole pattern of ocean temperature anomalies that extend down to the base of the deep winter mixed layer. These anomalies persist at depth through spring and summer within the stably stratified seasonal thermocline, insulated from the atmosphere by the formation of a shallow mixed layer in response to increasing solar radiation and weakening stirring due to slackened surface winds. Their sequestration ends in the following fall or early winter when the mixed layer deepens again due to the seasonal intensification of the extratropical atmospheric circulation, and the thermal anomalies created the previous winter become re-entrained into the mixed layer, affecting SST. This re-entrainment thus leads to the “reemergence” of the previous winter’s SST anomalies.

The oceanic response to NAO variability is also evident in changes in the distribution and intensity of winter convective activity in the northern North Atlantic. The convective renewal of intermediate and deep waters in the Labrador Sea and the Greenland–Iceland–Norwegian (GIN) Seas contribute significantly to the production and export of North Atlantic Deep Water and, thus, help to drive the global thermohaline circulation. The intensity of winter convection at these sites is not only characterized by large interannual variability, but also interdecadal variations that appear to be synchronized with variations in the NAO (Dickson et al., 1996). Deep convection over the Labrador Sea, for instance, was at its weakest and shallowest in the postwar instrumental record during the late 1960s. Since then, Labrador Sea Water has become progressively colder and fresher, with intense convective activity to unprecedented ocean depths (>2300 m) in the early 1990s (Visbeck et al., 2003; their Fig. 10). In contrast, warmer and saltier deep waters in recent years are the result of suppressed convection in the GIN Seas, whereas tracer evidence suggests that intense convection likely occurred during the late 1960s (Schlosser et al., 1991). A recent comprehensive review of the impact of the NAO on the circulation and hydrography of the Nordic Seas is given in Furevik and Nilsen (2007).

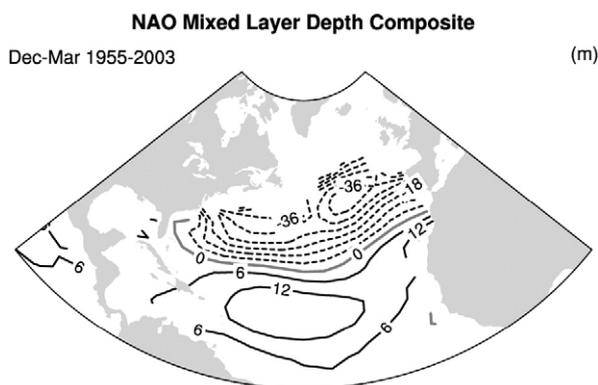


Fig. 15. Difference in mean winter (December–March) ocean mixed layer depth (m) between years when the NAO index (defined as in the middle panel of Fig. 10) exceeds one standard deviation over 1955–2003. The contour increment is 6 m and positive (negative) differences are given by the solid (dashed) contours.

5.4. Sea ice

The leading pattern of variability of winter Arctic sea ice concentrations exhibits a seesaw in ice extent between the Labrador and Greenland Seas. Strong interannual variability is evident in the sea ice changes, as are longer-term fluctuations including a trend over the past 30 years of diminishing (increasing) ice concentration during boreal winter east (west) of Greenland. Associated with the sea ice fluctuations are large-scale changes in SLP that closely resemble the NAO (Deser et al., 2000).

When the NAO is in its positive index phase, the Labrador Sea ice boundary extends farther south while the Greenland Sea ice boundary is north of its climatological extent (not shown). This is qualitatively consistent with the notion that the atmosphere directly forces the sea ice anomalies, either dynamically via wind-driven ice drift anomalies, or thermodynamically through surface air temperature anomalies. The relationship between the NAO index and an index of the North Atlantic ice variations is strong, although that it does not hold for all individual winters (Deser et al., 2000; Hilmer and Jung, 2000; Lu and Greatbatch, 2002) illustrates the importance of the regional atmospheric circulation in forcing the extent of sea ice. In general, the winters over the past decade have witnessed a retreat of the ice edge throughout the Arctic, even during the recent winters of weak NAO-related atmospheric circulation anomalies.

6. Summary and challenges

In this paper we have presented a basic review of modal variability over the North Atlantic. We have focused, in particular, on the NAO, as it is the dominant mode of regional climate variability, and it has been shown in many studies (including several in this special issue) to have profound impacts on a variety of ecological processes and, consequently, patterns of species abundance and dynamics (Alheit and Bakun, this issue). The NAO, for instance, controls fluctuations in temperature and salinity, vertical mixing, circulation patterns and ice formation of the North Atlantic Ocean, which affects marine biology through both direct and indirect pathways (e.g., Drinkwater et al., 2003). It is therefore critical to understand the mechanisms that control and affect the NAO and its temporal evolution. There is ample evidence that most of the atmospheric circulation variability in the form of the NAO arises from the internal, nonlinear dynamics of the extratropical atmosphere. Interactions between the time-mean flow and synoptic-timescale transient eddies are the central governing dynamical mechanism. As such, the month-to-month and even year-to-year changes in the phase and amplitude of the NAO are largely unpredictable. But that external forces might nudge the atmosphere to assume a high or low NAO index value over a particular month or season is important: even a small amount of predictability could be useful considering the significant impact the NAO exerts on the climate and ecosystems of the NH, and a better understanding of how the NAO responds to external forcing is crucial to the current debate on climate variability and change.

A number of different mechanisms that could influence the detailed state of the NAO have been proposed. Within the atmosphere itself, changes in the rate and location of tropical heating have been shown to be one way to influence the atmospheric circulation over the North Atlantic and, in particular, the NAO. Tropical convection, in turn, is sensitive to the underlying SST distribution, which exhibits much more persistence than SST variability in middle latitudes. This might lead, therefore, to some predictability of the NAO phenomenon. An example of tropical forcing of the NAO based on atmospheric general circulation modeling experiments is presented in Hoerling et al. (2004). In this study, approximately half of the recent upward trend in the wintertime NAO from 1950 to 1999 is attributable to increased precipitation over the tropical Indian Ocean in response to warming of the underlying sea surface. Sutton and Hodso (2003) also find

evidence of tropical Indian Ocean forcing of the NAO on long time scales, but in contrast to Hoerling et al. (2004), they conclude that this effect is secondary to forcing from the North Atlantic itself. In addition, recent modeling work has shown that the atmospheric response to the re-emerging North Atlantic SST tri-pole resembles the phase of the NAO that created the SST tri-pole the previous winter, thereby modestly enhancing the winter-to-winter persistence of the NAO (e.g., Cassou et al., 2007). Clearly, the importance of tropical versus extratropical ocean–atmosphere interaction on the NAO has not yet been fully determined.

Interactions with the lower stratosphere are also possibly important. This mechanism is of interest because it might also explain how changes in atmospheric composition influence the NAO. For example, changes in ozone, greenhouse gas (GHG) concentrations and/or levels of solar output affect the radiative balance of the stratosphere that, in turn, modulates the strength of the winter polar vortex. Given the relatively long time scales of stratospheric circulation variability (anomalies persist for weeks) dynamic coupling between the stratosphere and the troposphere via wave mean flow interactions could yield a useful level of predictive skill for the wintertime NAO. One recent example of the role of the lower stratosphere in the wintertime NAO involves Eurasian snow cover conditions. Cohen et al. (2002) and Gong et al. (2002) detail a mechanism whereby autumnal snow extent over Eurasia influences the winter NAO via dynamical coupling between the stratosphere and troposphere, with above (below) normal October snow extent leading to a negative (positive) phase of the winter NAO.

One of the most urgent challenges is to advance our understanding of the interaction between GHG forcing and the NAO. It now appears as though there may well be a deterministic relationship, which might allow for moderate low frequency predictability and thus needs to be studied carefully. Also, while the predictability of seasonal to interannual NAO variability will most likely remain low, some applications may benefit from the fact that this phenomenon leaves long-lasting imprints on surface conditions, in particular over the oceans. At the same time, the response of marine and terrestrial ecosystems to a shift in the NAO index might enhance or reduce the atmospheric carbon dioxide levels and thus provide a positive or negative feedback.

Acknowledgements

Much of the research discussed in this paper was conducted with support from the U.S. CLIVAR Atlantic program of NOAA OGP. The authors also wish to thank Adam Phillips for producing the figures, Christophe Cassou for conducting the cluster analysis, and Arthur Miller, Tore Furevik, and Ken Drinkwater for their helpful reviews of an earlier version of the manuscript.

References

- Alexander, M.A., Deser, C., 1995. A mechanism for the recurrence of wintertime midlatitude SST anomalies. *J. Phys. Oceanogr.* 25, 122–137.
- Alexanderson, H., Schmidth, T., Iden, K., Tuomenvirta, H., 1998. Long-term variations of the storm climate over NW Europe. *Glob. Ocean Atmos. Sys.* 6, 97–120.
- Alheit, J., Bakun, A., 2009. Population synchronies within and between ocean basins: apparent teleconnections and implications as to physical–biological linkage mechanisms (this issue).
- Bacon, S., Carter, D.J.T., 1993. A connection between mean wave height and atmospheric pressure gradient in the North Atlantic. *Int. J. Climatol.* 13, 423–436.
- Barnston, A.G., Livezey, R.E., 1987. Classification, seasonality and persistence of low frequency atmospheric circulation patterns. *Mon. Weather Rev.* 115, 1083–1126.
- Battisti, D.S., Bhatt, U.S., Alexander, M.A., 1995. A modeling study of the interannual variability in the wintertime North Atlantic Ocean. *J. Climate* 8, 3067–3083.
- Bjerknes, J., 1964. Atlantic air–sea interaction. *Adv. Geophys.* 10, 1–82.
- Carter, D.J.T., 1999. Variability and trends in the wave climate of the North Atlantic, a review. *Proc. 9th ISOPE Conf., Brest*, vol. III, pp. 12–18.
- Cassou, C., Terray, L., 2001a. Oceanic forcing of the wintertime low frequency atmospheric variability in the North Atlantic European sector, a study with the ARPEGE model. *J. Climate* 14, 4266–4291.
- Cassou, C., Terray, L., 2001b. Dual influence of Atlantic and Pacific SST anomalies on the North Atlantic/Europe winter climate. *Geophys. Res. Lett.* 28, 3195–3198.
- Cassou, C., Terray, L., Hurrell, J.W., Deser, C., 2004a. North Atlantic winter climate regimes: spatial asymmetry, stationarity with time, and oceanic forcing. *J. Climate* 17, 1055–1068.
- Cassou, C., Deser, C., Terray, L., Hurrell, J.W., Drevillon, M., 2004b. Summer sea surface temperature conditions in the North Atlantic and their impact upon the atmospheric circulation in early winter. *J. Climate* 17, 3349–3363.
- Cassou, C., Deser, C., Alexander, M.A., 2007. Investigating the impact of reemerging sea surface temperature anomalies on the winter atmospheric circulation over the North Atlantic. *J. Climate* 20 (14), 3510–3526.
- Cayan, D.R., 1992a. Latent and sensible heat flux anomalies over the northern oceans, driving the sea surface temperature. *J. Phys. Oceanogr.* 22, 859–881.
- Cayan, D.R., 1992b. Latent and sensible heat flux anomalies over the northern oceans, the connection to monthly atmospheric circulation. *J. Climate* 5, 354–369.
- Cohen, J., Salstein, D., Saito, K., 2002. A dynamical framework to understand and predict the major Northern Hemisphere mode. *Geophys. Res. Lett.* 29, 1412. doi:10.1029/2001GL014117.
- Corti, S., Molteni, F., Palmer, T.N., 1999. Signature of recent climate change in frequencies of natural atmospheric circulation regimes. *Nature* 398, 799–802.
- Curry, R.G., McCartney, M.S., 2001. Ocean gyre circulation changes associated with the North Atlantic Oscillation. *J. Phys. Oceanogr.* 31, 3374–3400.
- Czaja, A., Robertson, A.W., Huck, T., 2003. The role of Atlantic ocean–atmosphere coupling in affecting North Atlantic Oscillation variability. In: Hurrell, J.W., Kushnir, Y., Ottersen, G., Visbeck, M. (Eds.), *The North Atlantic Oscillation, Climatic Significance and Environmental Impact*. AGU Geophys. Monogr., vol. 134, pp. 147–172.
- Dai, A., Fung, I.Y., Del Genio, A.D., 1997. Surface observed global land precipitation variations during 1900–88. *J. Climate* 10, 2943–2962.
- de Coëtlogon, G., Frankignoul, C., 2003. On the persistence of winter sea surface temperature in the North Atlantic. *J. Climate* 16, 1364–1377.
- Delworth, T.L., 1996. North Atlantic interannual variability in a coupled ocean–atmosphere model. *J. Climate* 9, 2356–2375.
- Deser, C., 2000. On the teleconnectivity of the Arctic Oscillation. *Geophys. Res. Lett.* 27, 779–782.
- Deser, C., Timlin, M.S., 1997. Atmosphere–ocean interaction on weekly time scales in the North Atlantic and Pacific. *J. Climate* 10, 393–408.
- Deser, C., Walsh, J.E., Timlin, M.S., 2000. Arctic sea ice variability in the context of recent atmospheric circulation trends. *J. Climate* 13, 617–633.
- Deser, C., Alexander, M.A., Timlin, M.S., 2003. Understanding the persistence of sea surface temperature anomalies in midlatitudes. *J. Climate* 16, 57–72.
- Dickson, R.R., Lazier, J., Meincke, J., Rhines, P., Swift, J., 1996. Long-term co-ordinated changes in the convective activity of the North Atlantic. *Prog. Oceanogr.* 38, 241–295.
- Dickson, R.R., Osborn, T.J., Hurrell, J.W., Meincke, J., Blindheim, J., Ådlandsvik, B., Vigne, T., Alekseev, G., Maslowski, W., 2000. The Arctic Ocean response to the North Atlantic Oscillation. *J. Climate* 13, 2671–2696.
- Dommenget, D., Latif, M., 2002. A cautionary note on the interpretation of EOFs. *J. Climate* 15, 216–225.
- Drinkwater, K.F., Belgrano, A., Borja, A., Conversi, A., Edwards, M., Greene, C.H., Ottersen, G., Pershing, A.J., Walker, H., 2003. The response of marine ecosystems to climate variability associated with the North Atlantic Oscillation. In: Hurrell, J.W., Kushnir, Y., Ottersen, G., Visbeck, M. (Eds.), *The North Atlantic Oscillation, Climatic Significance and Environmental Impact*. AGU Geophys. Monogr., vol. 134, pp. 211–234.
- Esbensen, S.K., 1984. A comparison of intermonthly and interannual teleconnections in the 700 mb geopotential height field during the Northern Hemisphere winter. *Mon. Weather Rev.* 112, 2016–2032.
- Feldstein, S.B., 2000. Teleconnections and ENSO, the timescale, power spectra, and climate noise properties. *J. Climate* 13, 4430–4440.
- Feldstein, S.B., 2002. The recent trend and variance increase of the annular mode. *J. Climate* 15, 88–94.
- Furevik, T., Nilsen, J.E.O., 2007. Large-scale atmospheric circulation variability and its impacts on the Nordic seas ocean climate – a review. In: Drange, H., Dokken, T., Furevik, T., Gerdes, R., Berger, W. (Eds.), *The Nordic Seas, An Integrated Perspective*. AGU Geophys. Monogr., vol. 158, pp. 105–136.
- Gillett, N.P., 2003. Climate change and the North Atlantic Oscillation. In: Hurrell, J.W., Kushnir, Y., Ottersen, G., Visbeck, M. (Eds.), *The North Atlantic Oscillation, Climatic Significance and Environmental Impact*. AGU Geophys. Monogr., vol. 134, pp. 193–209.
- Gong, G., Entekhabi, D., Cohen, J., 2002. A large-ensemble model study of the wintertime AO/NAO and the role of interannual snow perturbations. *J. Climate* 15, 3488–3499.
- Hilmer, M., Jung, T., 2000. Evidence for a recent change in the link between the North Atlantic Oscillation and Arctic sea ice export. *Geophys. Res. Lett.* 27, 989–992.
- Hoerling, M.P., Hurrell, J.W., Xu, T., Bates, G.T., Phillips, A.S., 2004. Twentieth century North Atlantic climate change. Part II. Understanding the effect of Indian Ocean warming. *Clim. Dyn.* 23, 391–405.
- Hurrell, J.W., 1995. Decadal trends in the North Atlantic Oscillation, regional temperatures and precipitation. *Science* 269, 676–679.
- Hurrell, J.W., van Loon, H., 1997. Decadal variations in climate associated with the North Atlantic oscillation. *Clim. Change* 36, 301–326.
- Hurrell, J.W., Kushnir, Y., Visbeck, M., Ottersen, G., 2003. An overview of the North Atlantic Oscillation. In: Hurrell, J.W., Kushnir, Y., Ottersen, G., Visbeck, M. (Eds.), *The North Atlantic Oscillation, Climatic Significance and Environmental Impact*. AGU Geophysical Monograph, vol. 134, pp. 1–35.
- 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., Osborn, T.J., Briffa, K.R., 2003. Pressure-based measures of the North Atlantic Oscillation (NAO), a comparison and an assessment of changes in the strength of the NAO and in its influence on surface climate parameters. In: Hurrell, J.W., Kushnir, Y., Ottersen, G., Visbeck, M. (Eds.), *The North Atlantic Oscillation, Climatic Significance and Environmental Impact*. AGU Geophys. Monogr., vol. 134, pp. 51–62.
- Kalnay, E., et al., 1996. The NCEP/NCAR 40-year reanalysis project. *Bull. Am. Meteorol. Soc.* 77, 437–471.
- Kushnir, Y., Wallace, J.M., 1989. Low frequency variability in the Northern Hemisphere winter, geographical distribution, structure and time dependence. *J. Atmos. Sci.* 46, 3122–3142.
- Kushnir, Y., Cardone, V.J., Greenwood, J.G., Cane, M., 1997. On the recent increase in North Atlantic wave heights. *J. Climate* 10, 2107–2113.
- Kutzbach, J.E., 1970. Large-scale features of monthly mean Northern Hemisphere anomaly maps of sea-level pressure. *Mon. Weather Rev.* 98, 708–716.
- Lorenz, E.N., 1951. Seasonal and irregular variations of the Northern Hemisphere sea-level pressure profile. *J. Meteorol.* 8, 52–59.
- Lorenz, E.N., 1963. Deterministic nonperiodic flow. *J. Atmos. Sci.* 20, 130–141.
- Lu, J., Greatbatch, R.J., 2002. The changing relationship between the NAO and Northern Hemisphere climate variability. *Geophys. Res. Lett.* 29. doi:10.1029/2001GL014052.
- Luterbacher, J., et al., 2002. Extending North Atlantic Oscillation reconstructions back to 1500. *Atmos. Sci. Lett.* 2, 114–124. doi:10.1006/asle.2001.0044.
- Marshall, J., Kushnir, Y., Battisti, D., Chang, P., Czaja, A., Dickson, R., Hurrell, J., McCartney, M., Saravanan, R., Visbeck, M., 2001a. North Atlantic climate variability, phenomena, impacts and mechanisms. *Int. J. Climatol.* 21, 1863–1898.
- Marshall, J., Johnson, H., Goodman, J., 2001b. A study of the interaction of the North Atlantic Oscillation with ocean circulation. *J. Climate* 14, 1399–1421.
- Michelangeli, P.A., Vautard, R., Legras, B., 1995. Weather regime occurrence and quasi stationarity. *J. Atmos. Sci.* 52, 1237–1256.
- Miller, K.A., Munro, G.R., 2004. Climate and cooperation: a new perspective on the management of shared fish stocks. *Mar. Resour. Econ.* 19 (3), 367–393.
- Monahan, A.H., Fyfe, J.C., Flato, G.M., 2000. A regime view of northern hemisphere atmospheric variability and change under global warming. *Geophys. Res. Lett.* 27, 1139–1142.
- Monahan, A.H., Pandolfo, L., Fyfe, J.C., 2001. The preferred structure of variability of the Northern Hemisphere atmospheric circulation. *Geophys. Res. Lett.* 28, 1019–1022.
- Mysterud, A., Stenseth, N.C., Yoccoz, N.G., Ottersen, G.O., Langvatn, R., 2003. The response of terrestrial ecosystems to climate variability associated with the North Atlantic Oscillation. In: Hurrell, J.W., Kushnir, Y., Ottersen, G., Visbeck, M. (Eds.), *The North Atlantic Oscillation, Climatic Significance and Environmental Impact*. AGU Geophys. Monogr., vol. 134, pp. 235–262.
- North, G.R., Bell, T.L., Cahalan, R.F., Moeng, F.J., 1982. Sampling errors in the estimation of empirical orthogonal functions. *Mon. Weather Rev.* 110, 699–706.
- Palmer, T.N., 1999. A nonlinear dynamical perspective on climate prediction. *J. Climate* 12, 575–591.
- Portis, D.H., Walsh, J.E., El Hamly, M., Lamb, P.J., 2001. Seasonality of the North Atlantic Oscillation. *J. Climate* 14, 2069–2078.
- Rodwell, M.J., 2003. On the predictability of North Atlantic climate. In: Hurrell, J.W., Kushnir, Y., Ottersen, G., Visbeck, M. (Eds.), *The North Atlantic Oscillation, Climatic Significance and Environmental Impact*. AGU Geophys. Monogr., vol. 134, pp. 173–192.
- Rogers, J.C., 1981. Spatial variability of seasonal sea-level pressure and 500-mb height anomalies. *Mon. Weather Rev.* 109, 2093–2106.
- Rogers, J.C., 1984. The association between the North Atlantic Oscillation and the Southern Oscillation in the Northern Hemisphere. *Mon. Weather Rev.* 112, 1999–2015.
- Rogers, J.C., 1990. Patterns of low-frequency monthly sea level pressure variability (1899–1986) and associated wave cyclone frequencies. *J. Climate* 3, 1364–1379.
- Rogers, J.C., 1997. North Atlantic storm track variability and its association to the North Atlantic Oscillation and climate variability of Northern Europe. *J. Climate* 10, 1635–1645.
- Schlosser, P., Bonisch, G., Rhein, M., Bayer, R., 1991. Reduction of deepwater formation in the Greenland Sea during the 1980s, evidence from tracer data. *Science* 251, 1054–1056.
- Serreze, M.C., Carse, F., Barry, R.G., Rogers, J.C., 1997. Icelandic low activity, climatological features, linkages with the NAO, and relationships with recent changes in the Northern Hemisphere circulation. *J. Climate* 10, 453–464.
- Slonosky, V.C., Yiou, P., 2001. Secular changes in the North Atlantic Oscillation and its influence on 20th century warming. *Geophys. Res. Lett.* 28, 807–810.
- Stenseth, N.C., Ottersen, G., Hurrell, J.W., Mysterud, A., Lima, M., Chan, K.-S., Yoccoz, N.G., Adlandsvik, B., 2003. Studying climate effects on ecology through the use of climate indices, the North Atlantic Oscillation, El Niño Southern Oscillation and beyond. *Proc. R. Soc. Lond., B Biol. Sci.* 270, 2087–2096.
- Stenseth, N.C., Ottersen, G., Hurrell, J.W., Belgrano, A., 2005. *Marine Ecosystems and Climate Variation*. Oxford University Press. 252 pp.
- Straile, D., Livingstone, D.M., Weyhenmeyer, G.A., George, D.G., 2003. The response of freshwater ecosystems to climate variability associated with the North Atlantic Oscillation. In: Hurrell, J.W., Kushnir, Y., Ottersen, G., Visbeck, M. (Eds.), *The North Atlantic Oscillation, Climatic Significance and Environmental Impact*. AGU Geophysical Monograph, vol. 134, pp. 263–279.
- Sutton, R.T., Hodso, D.L.R., 2003. The influence of the ocean on North Atlantic climate variability 1871–1999. *J. Climate* 16, 3296–3313.
- 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.
- Thompson, D.W.J., Wallace, J.M., 2000. Annular modes in the extratropical circulation. Part I. Month-to-month variability. *J. Climate* 13, 1000–1016.
- Thompson, D.W.J., Lee, S., Baldwin, M.P., 2003. Atmospheric processes governing the Northern Hemisphere Annular Mode/North Atlantic Oscillation. In: Hurrell, J.W., Kushnir, Y., Ottersen, G., Visbeck, M. (Eds.), *The North Atlantic Oscillation, Climatic Significance and Environmental Impact*. AGU Geophys. Monogr., vol. 134, pp. 81–112.
- Timlin, M.S., Alexander, M.A., Deser, C., 2002. On the reemergence of North Atlantic SST anomalies. *J. Climate* 15, 2707–2712.
- Trenberth, K.E., 1984. Signal versus noise in the Southern Oscillation. *Mon. Weather Rev.* 112, 326–332.
- Trenberth, K.E., Paolino, D.A., 1980. The Northern Hemisphere sea level pressure data set, trends, errors and discontinuities. *Mon. Weather Rev.* 108, 855–872.
- Trenberth, K.E., Paolino, D.A., 1981. Characteristic patterns of variability of sea level pressure in the Northern Hemisphere. *Mon. Weather Rev.* 109, 1169–1189.
- Trenberth, K.E., Branstator, G.W., Karoly, D., Kumar, A., Lau, N.-C., Ropelewski, C., 1998. Progress during TOGA in understanding and modeling global teleconnections associated with tropical sea surface temperatures. *J. Geophys. Res.* 103, 14291–14324.
- Ulbrich, U., Christoph, M., 1999. A shift of the NAO and increasing storm track activity over Europe due to anthropogenic greenhouse gas forcing. *Clim. Dyn.* 15, 551–559.
- Visbeck, M., Chassignet, E.P., Curry, R.G., Delworth, T.L., Dickson, R.R., Krahnmann, G., 2003. The ocean's response to North Atlantic Oscillation variability. In: Hurrell, J.W., Kushnir, Y., Ottersen, G., Visbeck, M. (Eds.), *The North Atlantic Oscillation, Climatic Significance and Environmental Impact*. AGU Geophysical Monograph, vol. 134, pp. 113–146.
- 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., Zhang, Y., Lau, K.-H., 1993. Structure and seasonality of interannual and interdecadal variability of geopotential height and temperature fields in the Northern Hemisphere troposphere. *J. Climate* 6, 2063–2082.
- Wang, G., Schimel, D., 2003. Climate change, climate modes, and climate impacts. *Ann. Rev. Environ. Resour.* 28, 1–28.
- Watanabe, M., Kimoto, M., 2000. On the persistence of decadal SST anomalies in the North Atlantic. *J. Climate* 13, 3017–3028.
- White, W.B., 1995. Design of a global observing system for gyre-scale upper ocean temperature variability. *Prog. Oceanogr.* 36, 169–217.