An Overview of the North Atlantic Oscillation

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The North Atlantic Oscillation (NAO) is one of the most prominent and recurrent patterns of atmospheric circulation variability. It dictates climate variability from the eastern seaboard of the United States to Siberia and from the Arctic to the subtropical Atlantic, especially during boreal winter, so variations in the NAO are important to society and for the environment. Understanding the processes that govern this variability is, therefore, of high priority, especially in the context of global climate change. This review, aimed at a scientifically diverse audience, provides general background material for the other chapters in the monograph, and it synthesizes some of their central points. It begins with a description of the spatial structure of climate and climate variability, including how the NAO relates to other prominent patterns of atmospheric circulation variability. There is no unique way to define the spatial structure of the NAO, or thus its temporal evolution, but several common approaches are illustrated. The relationship between the NAO and variations in surface temperature, storms and precipitation, and thus the economy, as well as the ocean and ecosystem responses to NAO variability, are described. Although the NAO is a mode of variability internal to the atmosphere, indices of it exhibit decadal variability and trends. That not all of its variability can be attributed to intraseasonal stochastic atmospheric processes points to a role for external forcings and, perhaps, a small but useful amount of predictability. The surface, stratospheric and anthropogenic processes that may influence the phase and amplitude of the NAO are reviewed.

1. INTRODUCTION

Over the middle and high latitudes of the Northern Hemisphere (NH), especially during the cold season months (November-April), the most prominent and recurrent pattern of atmospheric variability 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 produce large changes in the mean wind speed and direction over the Atlantic, the heat and moisture transport between the Atlantic and the neighboring continents, and the intensity and number of storms, their paths, and their weather. Agricultural harvests, water management, energy supply and demand, and yields from fisheries, among many other things, are directly affected by the NAO. Yet, despite this pronounced influence, many open issues remain about which climate processes govern NAO variability, how the phenomenon has varied in the past or will vary in the future, and whether it is at all predictable. These and other topics are dealt with in detail in the following chapters. Our intent is to provide general background material for these chapters, as well as synthesize some of the central points made by other authors.

The NAO is one of the oldest known world weather patterns, as some of the earliest descriptions of it were from seafaring Scandinavians several centuries ago. The history of scientific research on the NAO is rich, and *Stephenson et al.* [this volume] present a stimulating account of the major scientific landmarks of NAO research through time. They also note that, today, there is considerable renewed interest

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in the phenomenon. The NAO and its time dependence, for instance, appear central to the global change debate. Surface temperatures over the NH are likely warmer now than at any other time over the past millennium [*Mann et al.*, 1999; *Jones et al.*, 2001], and the rate of warming has been especially high (~ 0.15°C decade⁻¹) over the past 40 years or so [*Folland et al.*, 2001; *Hansen et al.*, 2002]. A substantial fraction of this most recent warming is linked to the behavior of the NAO [*Hurrell*, 1996; *Thompson et al.*, 2000; also section 5.1], in particular a trend in its index from large amplitude anomalies of one phase in the 1960s to large amplitude anomalies of the opposite phase since the early 1980s. This change in the atmospheric circulation of the North Atlantic accounts for several other remarkable alterations in weather and climate over the extratropical NH as well, and it has added considerably to the debate over our ability to detect and distinguish between natural and anthropogenic climate change. Improved understanding of the relationship between the NAO and anthropogenic climate change has emerged as a key goal of modern climate research [*Gillett et al.*, this volume]. It has also made it critical to better understand how the NAO and its influence on surface climate has varied naturally in the past, either as measured from long instrumental records [*Jones et al.*, this volume] or estimated through multi-century multi-proxy reconstructions [*Cook*, this volume].

While it has long been recognized that the North Atlantic Ocean varies appreciably with the overlying atmosphere [*Bjerknes*, 1964], another reason for invigorated interest in the NAO is that the richly complex and differential responses of the surface-, intermediate- and deep-layers of the ocean to NAO forcing are becoming better documented and understood [*Visbeck et al.*, this volume]. The intensity of wintertime convective renewal of intermediate and deep waters in the Labrador Sea and the Greenland-Iceland-Norwegian (GIN) Seas, for instance, is not only characterized by large interannual variability, but also by interdecadal variations that appear to be synchronized with fluctuations in the NAO [e.g., *Dickson et al.*, 1996]. These changes in turn affect the strength and character of the Atlantic thermohaline circulation (THC) and the horizontal flow of the upper ocean, thereby altering the oceanic poleward heat transport and the distribution of sea surface temperature (SST).

On seasonal time scales, the upper North Atlantic Ocean varies primarily in response to changes in the surface winds, air-sea heat exchanges and freshwater fluxes associated with NAO variations [*Cayan*, 1992a,b]. This does not mean, however, that the extratropical interaction is only one-way. The dominant influence of the ocean on the overlying atmosphere is to reduce the thermal damping of atmospheric variations, and this influence becomes greater on longer time scales. The extent to which the influence of the ocean extends beyond this local thermodynamic coupling to affect the evolution and dynamical properties of the atmospheric flow is probably small, but the effect is non-zero [*Robinson*, 2000; *Kushnir et al.*, 2002]. The role of ocean-atmosphere coupling in determining the overall variability of the NAO is, therefore, a topic of much interest and ongoing research [*Czaja et al.*, this volume].

That the ocean may play an active role in determining the evolution of the NAO is also one pathway by which some limited predictability might exist [*Rodwell*, this volume]. New statistical analyses have revealed patterns in North Atlantic SSTs that precede specific phases of the NAO by up to 9 months, a link that likely involves the remarkable tendency of the extratropical ocean to preserve its thermal state throughout the year [*Kushnir et al.*, 2002]. On longer time scales, recent modeling evidence suggests that the NAO responds to slow changes in global ocean temperatures, with changes in the equatorial regions playing a central role [*Hoerling et al.*, 2001].

A second pathway that offers hope for improved predictability of the NAO involves links through which changes in stratospheric wind patterns might exert some downward control on surface climate [*Thompson et al.*, this volume]. A statistical connection between the month-to-month variability of the NH stratospheric polar vortex and the tropospheric NAO was established several years ago [e.g., *Perlwitz and Graf*, 1995], and more recently it has been documented that large amplitude anomalies in the wintertime stratospheric winds precede anomalous behavior of the NAO by 1-2 weeks [*Baldwin and Dunkerton*, 2001], perhaps providing some useful extended-range predictability. The mechanisms are not entirely clear, but likely involve the effect of the stratospheric flow on the refraction of planetary waves dispersing upwards from the troposphere [e.g., *Hartmann et al.*, 2000]. Similarly, processes that affect the stratospheric circulation on longer time scales, such as reductions in stratospheric ozone and increases in greenhouse gases, could factor into the trend in Atlantic surface climate observed over the past several decades [*Gillett et al.*, this volume]. Regardless of whether predictability arises from the influence of the ocean or from processes internal to the atmosphere, the salient point is that relatively little attention was paid to the NAO, until recently, because changes in its phase and amplitude from one winter to the next were considered unpredictable. The possibility that a small, but useful, percentage of NAO variance is predictable has motivated considerable recent research.

Finally, renewed interest in the NAO has come from the biological community. Variations in climate have a profound

influence on a variety of ecological processes and, consequently, patterns of species abundance and dynamics. Fluctuations in temperature and salinity, vertical mixing, circulation patterns and ice formation of the North Atlantic Ocean induced by variations in the NAO [*Visbeck et al.*, this volume] have a demonstrated influence on marine biology and fish stocks through both direct and indirect pathways [*Drinkwater et al.*, this volume]. This includes not only longer-term changes associated with interdecadal NAO variability, but interannual signals as well. Responses of terrestrial ecosystems to NAO fluctuations have also been documented [*Mysterud et al.*, this volume]. In parts of Europe, for example, many plant species have been blooming earlier and longer because of increasingly warm and wet winters, and variations in the NAO are also significantly correlated with the growth, development, fertility and demographic trends of many land animals. The NAO has a demonstrated influence on the physics, hydrology, chemistry and biology of freshwater ecosystems across the NH as well [*Straile et al.*, this volume]. Increasing awareness among and interactions between biologists and climate scientists will undoubtedly further our insights into the critical issue of the response of ecosystems to climate variability and climate change, and mutual interest in the NAO as a dominant source of climate variability is serving as an impetus for this interdisciplinary research.

For many reasons, then, there is broad and growing interest in the NAO. Improved understanding of the physical mechanisms that govern the NAO and its intraseasonal-to-interdecadal variability, and how modes of natural variability such as the NAO may be influenced by anthropogenic climate change, are research questions of critical importance. Setting the stage for the following more detailed review chapters, we begin with a description of the spatial structure of climate and climate variability, including a brief discussion of how the NAO is defined and how it relates to other, prominent patterns of atmospheric circulation variability. The impacts of the NAO on surface temperature, precipitation, storms, the underlying ocean and sea ice, and the local ecology are also briefly described, as are the mechanisms that most likely govern NAO variability. We conclude by expressing our thoughts on outstanding issues and future challenges.

2. THE SPATIAL STRUCTURE OF 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 struc-

ture 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 boreal winter (December-February) to boreal summer (June-August, Figure 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 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 meridional 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 (Figure 2). These middle latitude westerlies 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 (~ 5-6 km), the boreal winter map of the geopotential height field reveals a westward tilt with elevation of the high latitude surface cyclones and anticyclones (Figure 3). There is a clear "wavenumber two" configuration with low-pressure troughs over northeastern Canada and just east of Asia, and high-pressure ridges just to the west of Europe and 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 (Figure 3) are geographically anchored, they do change in time either because the heating patterns in the atmosphere vary or
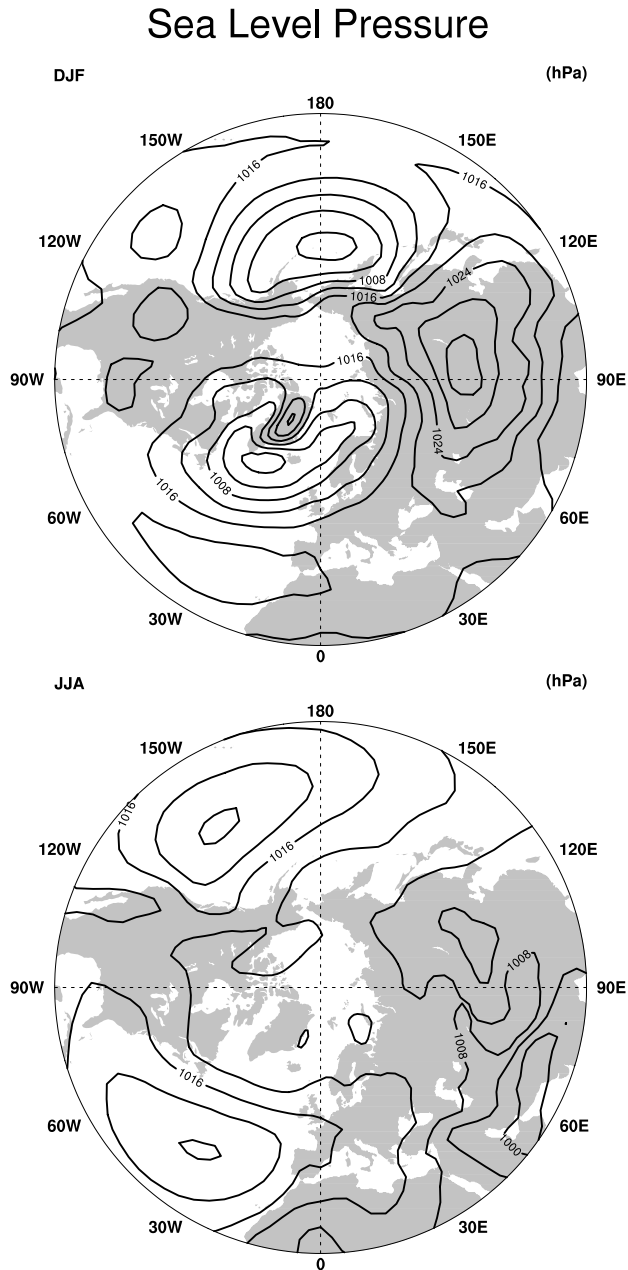
Sea Level Pressure

Figure 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 1958-2001 [*Kalnay et al.,* 1996], and the contour increment is 4 hPa.

because of internal (chaotic) processes. The amplitude and structure of the variability of the seasonal mean 500 hPa geopotential height field (Figure 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 (Figure 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 (Figure 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]. That these two patterns "stand out" above a background continuum comprised of a complete (hemispheric) set of one-point correlation maps is, of course, subjective [*Wallace,* 1996], but the strong consensus is that they do [e.g., *Kushnir and Wallace*, 1989].

The PNA teleconnection pattern has four centers of 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 (Figure 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
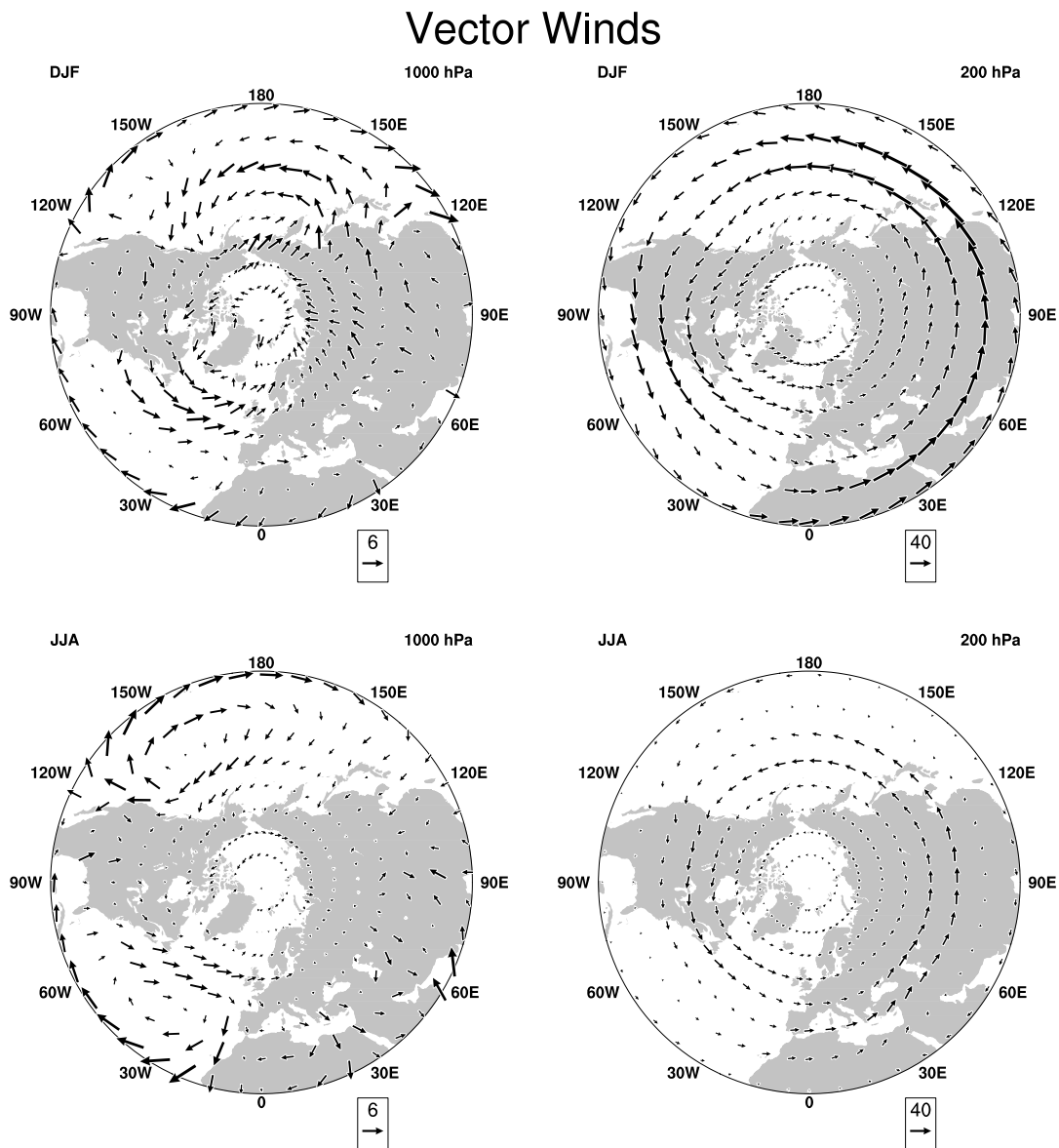
Vector Winds

Figure 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 1958-2001. The scaling vectors are indicated in the boxes and are given in units of m s^{-1}.

of the four centers of the PNA is their relation to the mean atmospheric circulation (Figure 3). As stated by *Kushnir* [2002], variations in the PNA pattern "represent variations in the waviness of the atmospheric flow in the western half-hemisphere and thus the 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, as is the similar Pacific South American (PSA) teleconnection pattern in the Southern Hemisphere [SH; *Kiladis and Mo,* 1998]. Significant variability of the PNA occurs

500hPa Geopotential Height
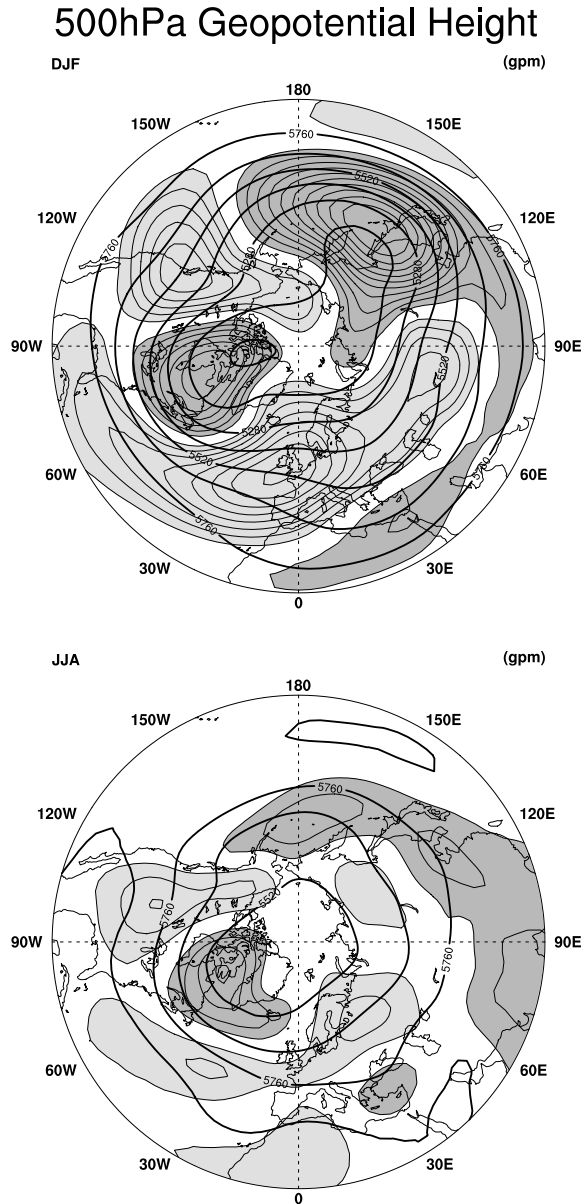
σ(z) 500hPa

Figure 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 gpm, over 1958-2001. The thin contours (every 20 gpm, zero contour excluded) indicate departures from the zonal average: negative (positive) departures are indicated by dark (light) shading.

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.*, this volume; *Czaja et al.*, this volume], as is evident from observations and climate model experiments that do not include
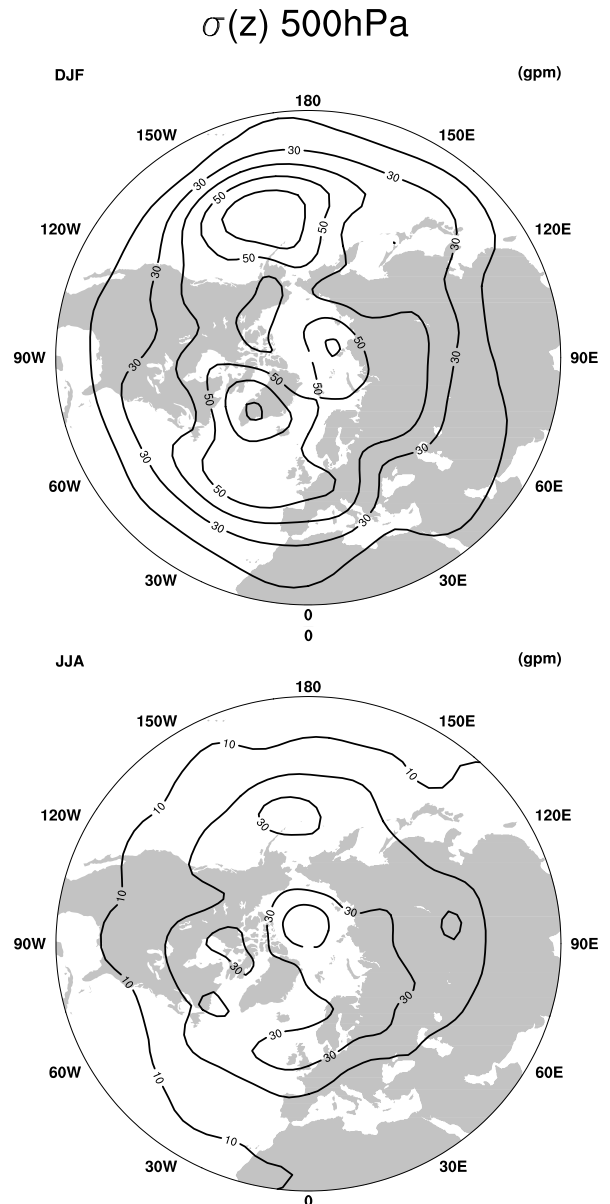
Figure 4. Interannual variability of 500 hPa geopotential height for (top) boreal winter (December-February) and (bottom) boreal summer (June-August) over 1958-2001. The contour increment is 10 gpm.

SST, sea ice or land surface variability (see section 6.1 and Figure 19). 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 (Figure 5; section 3). Both the NAO and PNA are also reflected in the spatial patterns of the two leading empirically-determined orthogonal functions (EOFs) of NH bore-

One-Point Correlation
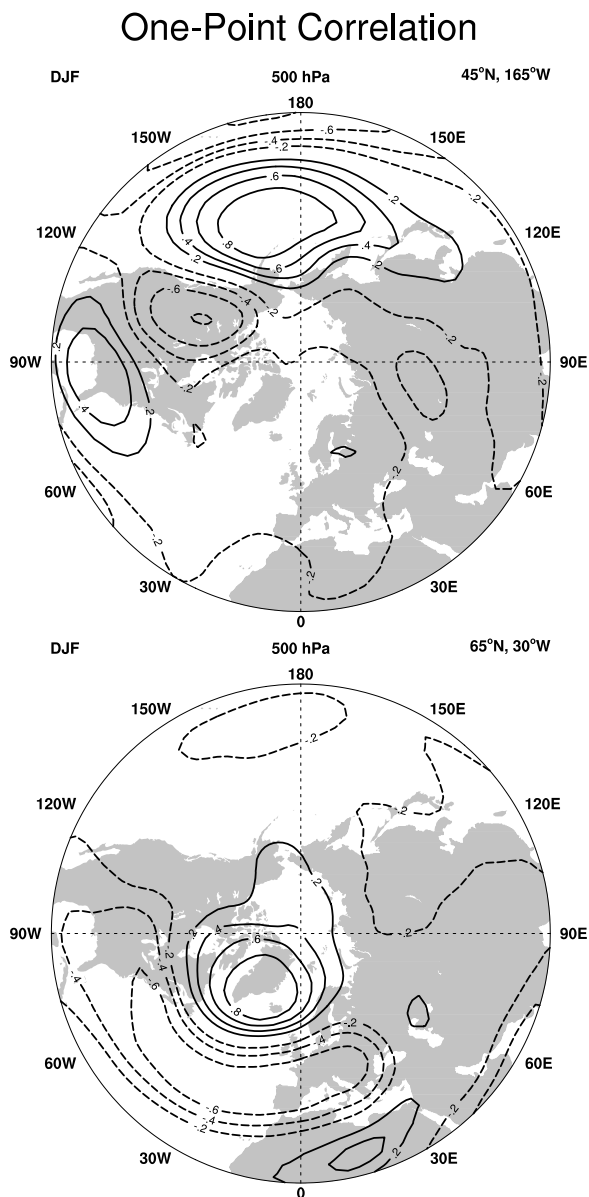
Figure 5. One-point correlation maps of 500 hPa geopotential heights for boreal winter (December-February) over 1958-2001. In the top panel, the reference point is 45ºN, 165ºW, corresponding to the primary center of action of the PNA pattern. In the lower panel, the NAO pattern is illustrated based on a reference point of 65ºN, 30ºW. Negative correlation coefficients are dashed, the contour increment is 0.2, and the zero contour has been excluded.

surface, however, where the NAO dominates the leading EOF of the NH SLP field [section 3.2; see also *Kutzbach,* 1970; *Rogers*, 1981; *Trenberth and Paolino,* 1981; *Thompson et al.*, this volume]. Analyzing SLP also 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.*, this volume]. Thus, in the following, we 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., Figure 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; see also *Monahan et al*., 2000; 2001]. 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 (20º-70ºN; 90ºW-40ºE) are illustrated in Figure 6. The patterns are very similar if based on the cross-correlation matrix (not shown). The patterns are

al winter 500 hPa height (not shown), but in order to see them clearly it is necessary to rotate (i.e., to form linear combinations of) the EOFs in a manner that tends to simplify their spatial structure [e.g., *Barnston and Livezey*, 1987; *Kushnir and Wallace*, 1989]. This is less of an issue at the
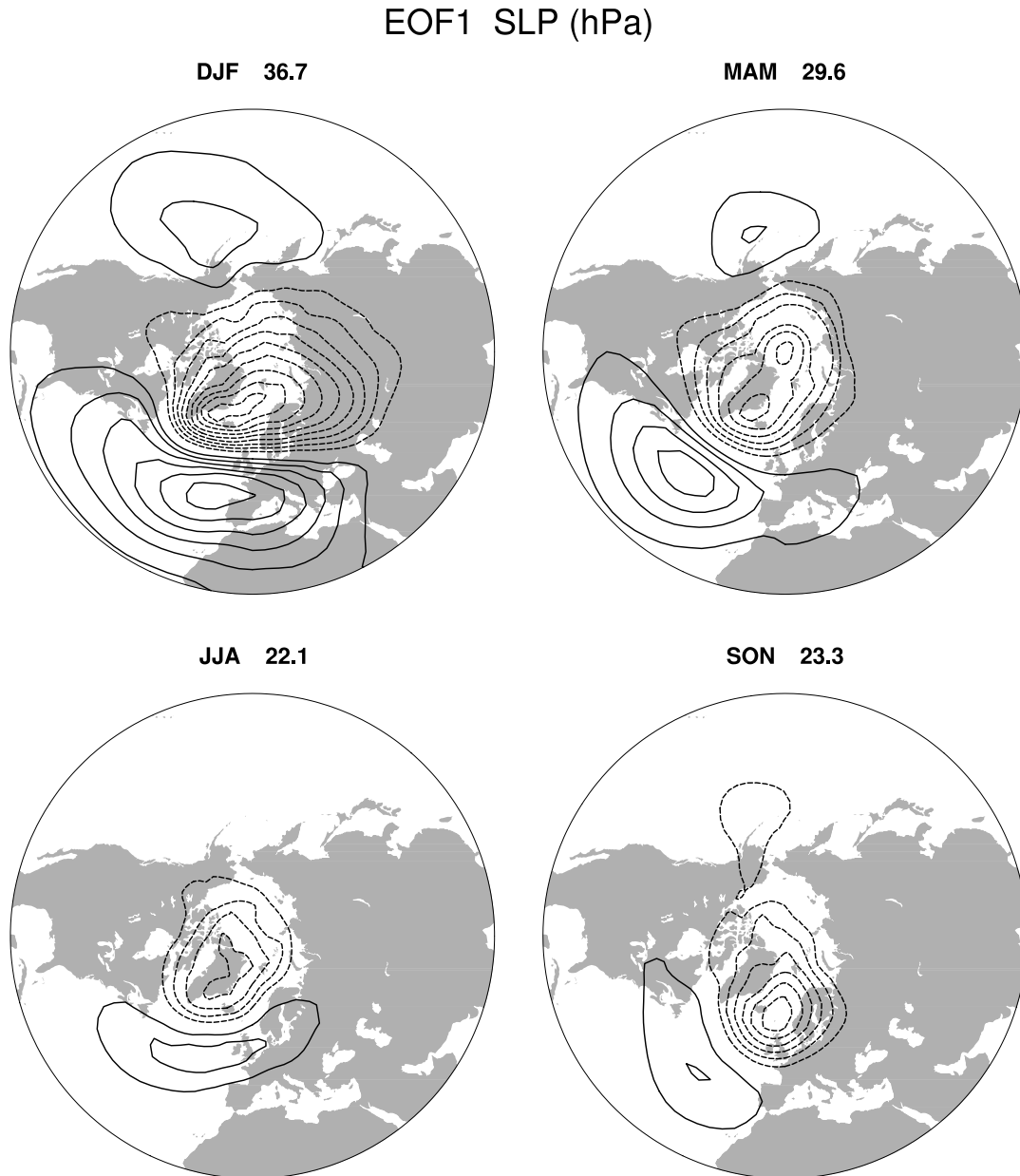
EOF1 SLP (hPa)

DJF   36.7   MAM   29.6

JJA   22.1   SON   23.3

Figure 6. Leading empirical orthogonal functions (EOF 1) of the seasonal mean sea level pressure anomalies in the North Atlantic sector (20º-70ºN, 90ºW-40ºE), 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 1899-2001 [see *Trenberth and Paolino,* 1980].

displayed in terms of amplitude, obtained by regressing the hemispheric SLP anomalies upon the leading principal component (PC) time series from the Atlantic domain.

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 "centers of action" 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 Figure 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 55ºN combine with a broad region of anomalously low pressure throughout the Arctic to enhance the climatological meridional pressure gradient (Figure 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 (Figure 7).

The NAO is well separated (and thus less likely to be affected by statistical sampling errors) in all seasons from
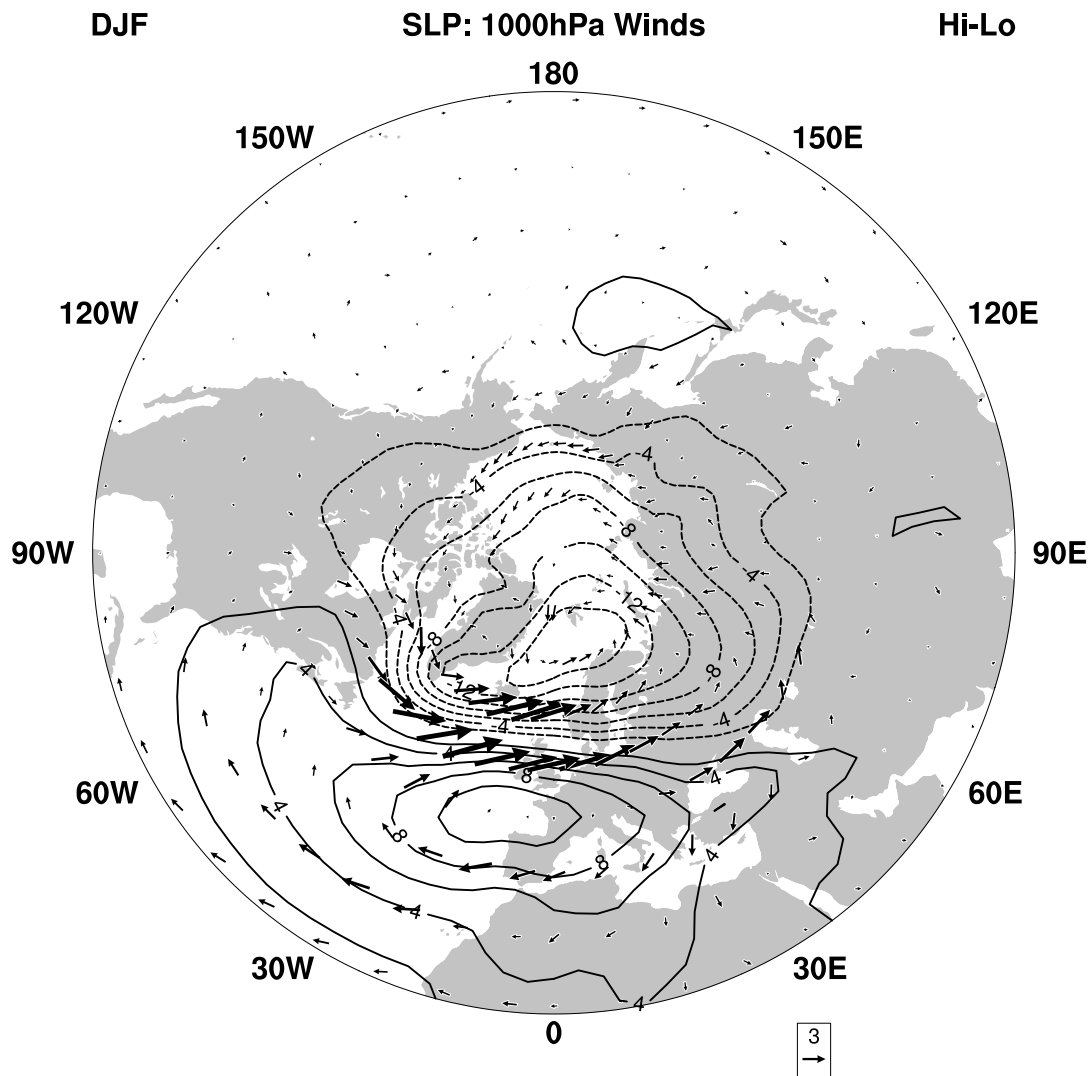
Figure 7. The difference in boreal winter (December-February) mean sea level pressure and 1000 hPa vector winds between positive (hi) and negative (lo) index phases of the NAO. The composites are constructed from winter data (the NCEP/NCAR reanalyses over 1958-2001) when the magnitude of the NAO index (defined as the principal component time series of the leading empirical orthogonal function of Atlantic-sector sea level pressure, as in Figures 6 and 10) exceeds one standard deviation. Nine winters are included in each composite. The contour increment for sea level pressure is 2 hPa, negative values are indicated by the dashed contours, and the zero contour has been excluded. The scaling vector is 3 m s^{-1}.

the second eigenvector, according to the criterion of *North et al.* [1982]. The second EOF, which resembles the East Atlantic (EA) pattern during the winter and spring months [*Wallace and Gutzler,* 1981; *Barnston and Livezey*, 1987], generally accounts for about 15% of the total SLP variance (not shown). 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.

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 (sections 5.1 and 5.2), as well as on ecosystems (section 5.4), is also greatest at this time of year. As most of the other chapters in this volume do as well, we focus hereafter on the winter variations. But that coherent fluctuations of surface pressure, temperature and precipitation occur throughout the year over the North Atlantic, and decadal and longer-term variability is not confined to winter, should not be lost on the reader. For instance, *Hurrell et al.* [2001; 2002] and *Hurrell and Folland* [2002] document significant interannual to multi-decadal fluctuations in the summer NAO pattern (Figure 6), including a trend toward persistent anticyclonic flow over northern Europe that has contributed to anomalously warm and dry conditions in recent decades [see also *Sexton et al.*, 2002; *Rodwell*, this volume]. 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 system [*Czaja et al.*, this volume; *Rodwell*, this volume]. Undoubtedly, further examinations of the annual cycle of climate and climate change over the Atlantic, as well as the mechanisms responsible for those variations, are needed.

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 (section 4), and its longer-term behavior reflects the combined effect of residence time in any given phase and its amplitude therein.

3.2. EOF Analysis of Northern Hemisphere 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., *Dommenget and Latif*, 2002]. These issues have been at the center of a recent debate [*Deser*, 2000; *Wallace*, 2000*; Ambaum et al.*, 2001] over whether or not the NAO is a regional expression of a larger-scale (hemispheric) mode of variability known as the Arctic Oscillation (AO) or, as it is more recently referred to, the NH Annular Mode [NAM; *Thompson et al.*, this volume].

The NAM is defined as the first EOF of NH (20°-90°N) winter SLP data (shown in Figure 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 (Figure 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 recently strongly 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 [see also *Wallace*, 2000; *Hartmann et al.*, 2000].

The arguments for the existence of the NAM, described in much more detail by *Thompson et al.* [this volume], include the following: (1) the zonally-symmetric component of the NAM is evident in the leading EOFs of heights and winds from the surface through the stratosphere, with variability in the latter region being dominated by a truly annular mode;

EOF1 SLP (hPa)

NAM   **DJFM**   **23.5**

NAO   **DJFM**   **39.6**
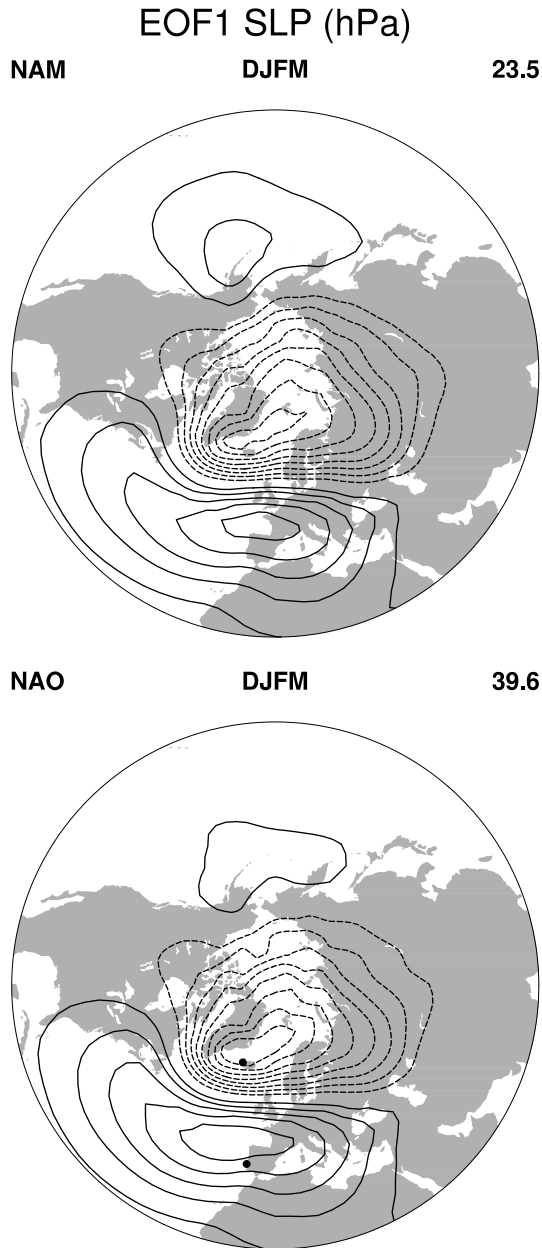
Figure 8. Leading empirical orthogonal function (EOF 1) of the winter (December-March) mean sea level pressure anomalies over (top) the Northern Hemisphere (20º-90ºN) and (bottom) the North Atlantic sector (20º-70ºN, 90ºW-40ºE), and the percentage of the total variance they explain. The patterns are 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 1899-2001 [see *Trenberth and Paolino,* 1980]. The dots in the bottom panel represent the locations of Lisbon, Portugal and Stykkisholmur, Iceland used in the station based NAO index of *Hurrell* [1995a] (see Figure 10).

(2) the strong similarity of the NAM to the spatial pattern of circulation variability in the SH, known as the Southern Annular Mode (SAM); (3) the "signature" of the NAM in the meridional profiles of the month-to-month variance of the zonally-averaged circulation; and (4) that the NAM seems to orchestrate weather and climate over the hemisphere, not just the Atlantic sector, on time scales from weeks to decades. This point of view clearly suggests that the NAM reflects dynamical processes that transcend the Atlantic sector. It is not a view that is universally accepted, however [*Kerr*, 1999].

Deser [2000] has argued that the NAM is not a teleconnection pattern in the sense that there are only weak correlations between the Atlantic and Pacific middle latitude centers on both intraseasonal (month-to-month) and interannual time scales. In addition, while interannual fluctuations in SLP over the Arctic and Atlantic centers of action are significantly (negatively) correlated (e.g., Figure 5), the Arctic and Pacific centers are not. This leads her to conclude "the annular character of the AO is more a reflection of the dominance of its Arctic center of action than any coordinated behavior of the Atlantic and Pacific centers". *Ambaum et al.* [2001] reach a similar conclusion, but also based on an assessment of the physical consistency between the NAM and NAO structures in SLP and the leading patterns of variability in other, independent climate variables. In particular, they show that leading EOFs of SLP, lower tropospheric winds and temperature over the Atlantic sector are dynamically related and are clear representations of the NAO, while the same analysis applied to the hemispheric domain yields very different results and patterns that are not obviously related. Rather, over the Pacific sector, they show that dynamical consistency among fields emerges for the PNA. *Ambaum et al.* [2001] also note that NAM variability is superposed upon a strongly zonally asymmetric climatology (Figures 1-3; note that the Icelandic and Aleutian low pressure centers occupy different latitudes), so that it does not correspond to a uniform modulation of the climatological features. In the positive NAM phase (depicted in Figure 8), the North Atlantic tropospheric subtropical and polar jets (Figure 2) are strengthened, but the subtropical jet in the Pacific is weakened.

While the above arguments suggest that the NAO paradigm may be more robust and physically relevant for NH variability, the debate is not over. Recently, for instance, *Wallace and Thompson* [2002] suggest that the lack of teleconnectivity between the Atlantic and Pacific sectors is consistent with the NAM if a second mode is present that favors out-of-phase behavior between these sectors. They suggest this mode could be the PNA. Regardless, the important point is that the physical mechanisms associated with annular

mode behavior may be very relevant to understanding the existence of the NAO, regardless of the robustness of the NAM paradigm. For instance, as previously noted, the leading wintertime pattern of variability in the lower stratosphere is clearly annular, 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., *Perlwitz and Graf*, 1995; *Kodera et al.*, 1996; *Thompson and Wallace*, 1998; *Deser*, 2000]. *Thompson et al.* [this volume] 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 nonlinear approaches, such as cluster analysis or non-linear principal component analysis [*Monahan et al.*, 2000; 2001]. Here we apply the former to 100 years of December-March monthly SLP data using the procedures of *Cassou and Terray* [2001a,b], 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).

The clustering algorithm applied over the Atlantic domain (20º-70ºN; 90ºW-40ºE) identifies four winter climate regimes in SLP (Figure 9). Two of them correspond to the negative and positive phases of the NAO, while the third and fourth regimes display a strong anticyclonic ridge and trough, respectively, off western Europe and bear some resemblance to the EA teleconnection pattern [*Wallace and Gutzler*, 1981; *Barnston and Livezey*, 1987]. Both the ridge and negative NAO regimes occur in about 30% of all winter months since 1900, while both the positive NAO and trough regimes occur in about 20% of all winter months. These numbers are sensitive to the period of analysis, reflecting that the dominance of certain regimes over others varies over time (section 4).

In contrast to the typical NAO pattern identified through linear approaches (e.g., Figures 5 and 6), some interesting spatial asymmetries are evident in Figure 9. Most striking is the difference in the position of the middle latitude pressure anomalies between the two NAO regimes: in particular, the eastward shift (by ~30º longitude) in the positive relative to the negative regime. The main difference in the northern center is the northeastward extension of SLP anomalies during positive NAO regime months. These spatial asymmetries are not dependent on the analysis period: they are evident in subperiods of the ~100-year long SLP data set [C.

Cassou, personal communication]. Similar results, indicating a non-linearity in NAO variability, are found when the PC time series of the leading EOF of Atlantic SLP (Figure 8) is used to define and average together positive and negative index winters (like those used to construct Figure 7).

The robustness of the eastward displacement of the NAO in positive regime months has interesting implications for conclusions drawn recently from climate model studies on how increasing greenhouse gas (GHG) concentrations might affect the spatial structure of the NAO [*Gillett et al.*, this volume]. *Ulbrich and Christoph* [1999], for instance, concluded that future enhanced GHG forcing 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 excitement 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. *Walker and Bliss* [1932] constructed the first index of the NAO using a linear combination of surface pressure and temperature measurements from weather stations on both sides of the Atlantic basin [see also *Wallace*, 2000; *Wanner et al.*, 2001; *Stephenson et al.*, this volume]. In the mid-20th century, indices of the "zonal index cycle" were popular [*Namias*, 1950; *Lorenz*, 1951 among others]. These indices characterize variations in the strength of the zonally averaged middle latitude surface westerlies and thus largely reflect variations in the NAO [*Wallace*, 2000; *Stephenson et al.*, this volume; *Thompson et al.*, this volume]. European scientists have introduced many others, all also strongly related to the NAO but generally not well known. *Stephenson et al.* [this volume] describe several of them. An example is the "westerly index" of *Lamb* [1972], one of several indices associated with a set of circulation types relevant to the climate of the United Kingdom that are still used in research today [C. Folland, personal communication].

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.
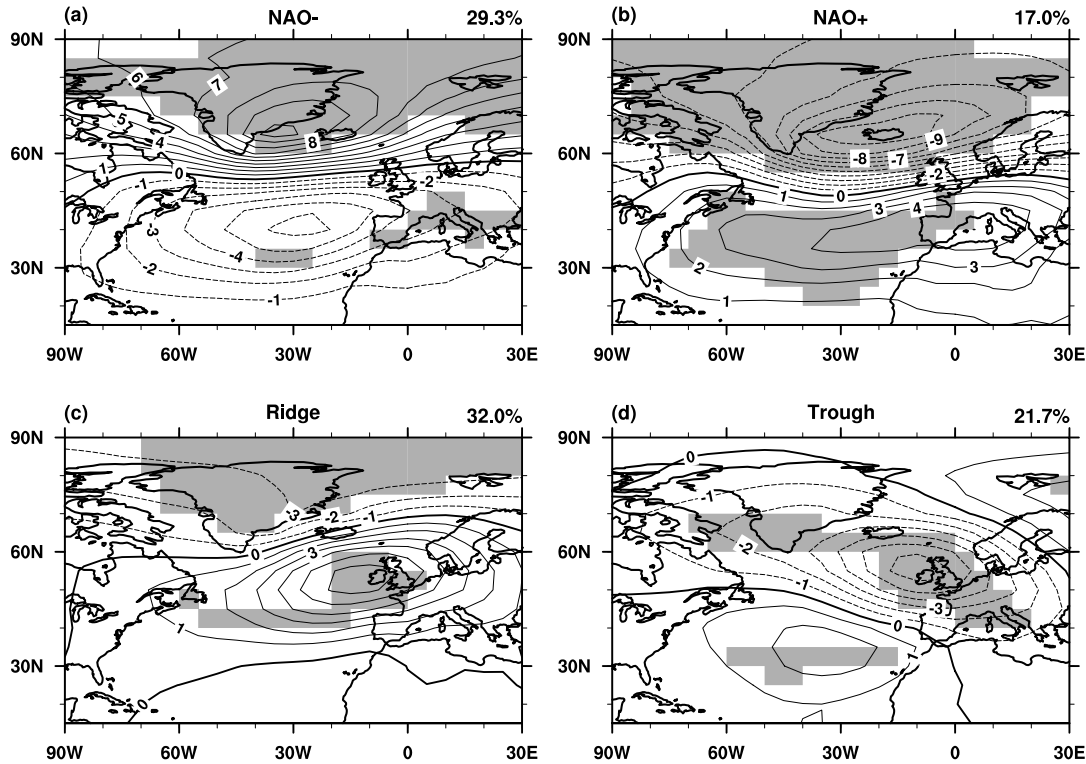
Figure 9. Boreal winter (December-March) climate regimes in sea level pressure (hPa) over the North Atlantic domain (20°-70°N, 90°W-40°E) using monthly data over 1900-2001. Shaded areas exceed the 95% confidence level using T and F statistics [see *Cassou*, 2001]. The percentage at the top right of each panel expresses the frequency of occurrence of a cluster out of all winter months since 1900. The contour interval is 1 hPa.

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*, 1995a; *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.* [this volume] discuss and compare various station-based indices in detail. They note that a major advantage of most of these indices is their extension back to the mid-19th century or earlier, and they even present a new instrumental NAO index from London and Paris records dating back to the late 17th century [see also *Slonosky et al.*, 2001].

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 (Figure 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.*, this volume]. 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 parts of the 20th century, 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. We also present the time history of occurrence of the NAO regimes identified in Figure 9. All comparisons are for the winter (December-March) season. *Osborn et al.* [1999], *Wallace* [2000], *Wanner et al.* [2001], *Portis et al.* [2001], and *Jones et al.* [this volume] present quantitative comparisons of these and other NAO-related indices, the latter two papers for other seasons as well.

4.1. Time Series

Rogers [1984] simplified the NAO index of *Walker and Bliss* [1932] by examining the difference in normalized SLP anomalies from Ponta Delgada, Azores and Akureyri, Iceland. Normalization is used to avoid the series being

dominated by the greater variability of the northern station (e.g., Figure 4). *Hurrell* [1995a] analyzed the important coupled modes of wintertime variability in SLP and surface temperature over the North Atlantic sector, and concluded that the southern-node station of Lisbon, Portugal better captured NAO-related variance (e.g., Figure 8). Using Lisbon also allowed him to extend the record a bit further back in time (to 1864), and *Jones et al.* [1997] subsequently showed that an adequate index could be obtained using the even longer record from Gibraltar (to 1821). *Jones et al.* [this volume] show that all of these indices are highly correlated on interannual and longer time scales, but that the choice of the southern station does make some difference. In contrast, the specific location of the northern node (among stations in Iceland) is not critical since the temporal variability over this region is much larger than the spatial variability. For instance, December-March anomalies in SLP at Stykkisholmur and Akureyri correlate at 0.98 [*Hurrell and van Loon*, 1997].

The winter-mean index of *Hurrell* [1995a] is shown in Figure 10. Positive values of the index indicate stronger-than-average westerlies over the middle latitudes associated with pressure anomalies of the like depicted in Figures 6 and 8. 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 1899-2002, indicating that the station-based index adequately represents the time variability of the winter-mean NAO spatial pattern. Moreover, it correlates with PC1 of NH SLP [the NAM index of *Thompson and Wallace*, 1998] 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 intraseasonal 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, but the correlations involving the station-based index remain unchanged.

One conclusion from Figure 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, and there is also a considerable amount of variability within a given winter season [*Nakamura*, 1996; *Feldstein*, 2000; see also Figure 11]. This is consistent with the notion that much of the atmospheric circulation variability in the form of the NAO arises from processes internal to the atmosphere [section 6.1; *Thompson et al.*, this volume], in which various scales of motion interact with one another to produce random (and thus unpredictable) variations. There are, however, periods when anomalous NAO-like circulation patterns persist over quite a few consecutive winters. In the subpolar

North Atlantic, for instance, SLP tended to be anomalously low during winter from the turn of the 20th century until about 1930 (positive NAO index), while the 1960s were characterized by unusually high surface pressure and severe winters from Greenland across northern Europe [negative NAO index; *van Loon and Williams*, 1976; *Moses et al.*, 1987]. A sharp reversal occurred from the minimum index values in the late 1960s to strongly positive NAO index values in the early and mid 1990s. Whether such low frequency (interdecadal) NAO variability arises from interactions of the North Atlantic atmosphere with other, more slowly varying components of the climate system such as the ocean [*Czaja et al.*, this volume; *Visbeck et al.*, this volume], whether the recent upward trend reflects a human influence on climate [*Gillett et al.*, this volume], or whether the longer time scale variations in the relatively short instrumental record simply reflect finite sampling of a purely random process [*Czaja et al.*, this volume] driven entirely by atmospheric dynamics [*Thompson et al.*, this volume] will be discussed further in section 6.

Another index, the time history of the occurrence of NAO regimes, offers a different perspective (Figure 11). Plotted is the number of months in any given winter that either or both of the NAO regimes (Figure 9) occur. 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 months are found during the 1960s, while very few negative regime months have been observed recently, consistent with the upward trend in the indices of Figure 10 and the "eastward shift" of the NAO centers of action in recent decades [Figure 9; *Hilmer and Jung*, 2000; *Lu and Greatbatch*, 2002].

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. Over the ~100-year record, for example, all four-winter months are classified as the negative NAO regime for only four winters (1936, 1940, 1969, and 1977, the year given by January), only two winters (1989 and 1990) have more than two months classified under the positive NAO regime, and there are nine winters during which neither NAO regime can be identified in any months. This is also a reminder of the second point: 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 by the canonical NAO pattern in Figures 6 and 8. One notable difference

SLP-based Indices (Dec-Mar)
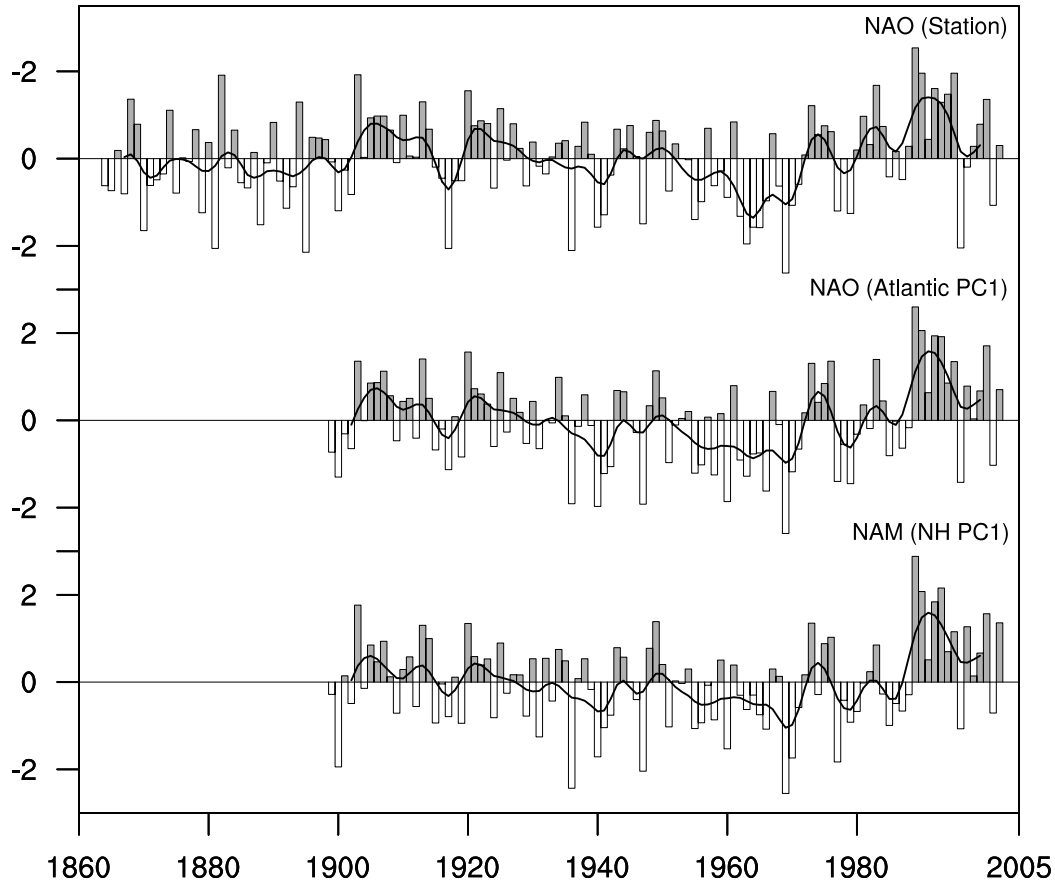
Figure 10. Normalized indices of the mean winter (December-March) NAO constructed from sea level pressure data. In the top panel, the index is based on the difference of normalized sea level pressure between Lisbon, Portugal and Stykkisholmur/Reykjavik, Iceland from 1864 through 2002. The average winter 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 panel, the index is the principal component time series of the leading empirical orthogonal function (EOF) of Atlantic-sector sea level pressure (bottom panel of Figure 8). In the lower panel, the index is the principal component time series of the leading EOF of Northern Hemisphere sea level pressure (top panel of Figure 8). 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). See http://www.cgd.ucar.edu/~jhurrell/nao.html for updated time series.

between the time history of NAO regime occurrences and more conventional NAO indices occurs early in the 20th century. All three conventional indices have generally positive values from 1900 until about 1930 (Figure 10). However, the ridge and trough regimes (Figure 9) were more dominant than the NAO regimes over this period (as can be deduced by the small occupancy rates in Figure 11; see *Cassou* [2001] for more discussion). This is also consistent with *van Loon and Madden* [1983], who show that the early 20th century was characterized by a southward displacement of the maximum North Atlantic SLP variance from the Irminger Sea to

near Ireland (their Figures 1 and 3). That the ridge regime projects upon the positive index phase of the NAO (Figure 9) helps explain the strong positive values of the conventional indices, but it also warns that reducing the complexities of the North Atlantic atmospheric circulation to one simple index can be misleading.

4.2. Power Spectrum

Spectral analysis is used to quantify periodicities in a time series and to gain insights into the dynamical processes
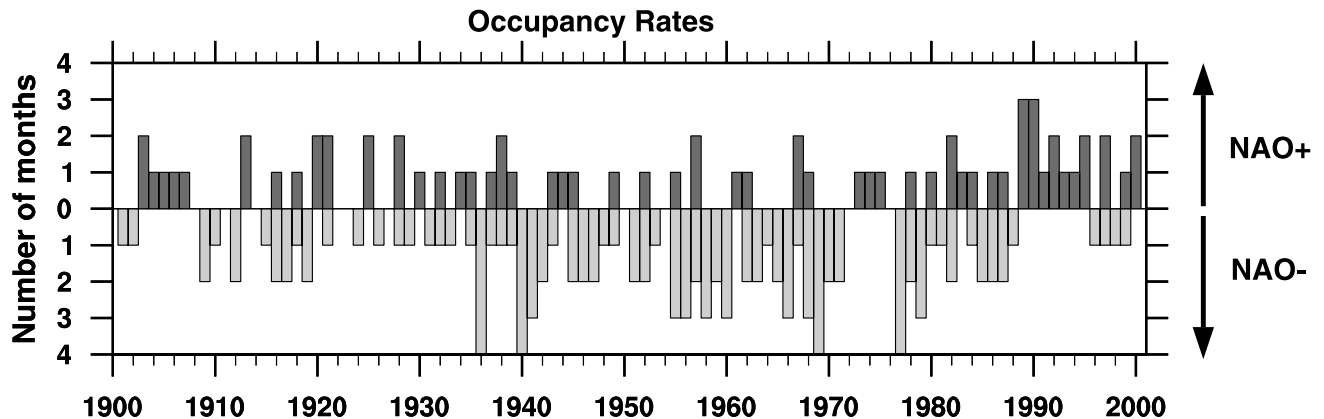
Figure 11. The time history of occurrence of the NAO regimes (panels a and b of Figure 9) over 1900-2000. The vertical bars give the number of months in each winter (December-March) season that the given regime is present. The indicated year corresponds to the January of the winter season (e.g., 1990 is the winter of 1989/1990).

associated with modes of climate variability. Numerous authors using different techniques have examined the power spectra of NAO indices, and *Greatbatch* [2000] and *Wanner et al.* [2001] review many of these efforts. As is largely evident from Figure 10, the major conclusion is that no preferred time scale of NAO variability is evident. This is consistent with Figure 12, which shows the power spectrum of the NAO index defined as the PC1 time series of North Atlantic SLP (Figure 10, middle panel). Nearly identical spectra are obtained for the NAM and the longer station-based NAO indices as well (not shown).

The spectrum of the winter-mean NAO index is slightly "red", with power increasing with period. It reveals somewhat enhanced variance at quasi-biennial periods, a deficit in power at 3 to 6 year periods, and slightly enhanced power in the 8-10 year band, but no significant peaks. The power evident at the lowest frequencies reflects the trends evident in Figure 10. *Hurrell and van Loon* [1997] examined the time evolution of these signals using a station-based index, and found that the quasi-biennial variance was enhanced in the late 19th and early 20th centuries, while the 8-10 year variance was enhanced over the latter half of the 20th century. They also noted similar behavior in the spectra of European surface temperature records over the same period of analysis. *Appenzeller et al.* [1998], *Higuchi et al.* [1999], and *Wanner et al.* [2001] among others have also noted that apparent "bumps" in the NAO spectra come and go over time.

Feldstein [2000] examined the spectral characteristics of the NAO using daily data, and concluded that its temporal evolution is largely 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 (Figure 12) could entirely be a remnant of the energetic weekly vari-

ability [see also *Wunsch*, 1999; *Stephenson et al.*, 2000]. In this "climate noise paradigm" [*Leith*, 1973; *Madden*, 1976], NAO variability is entirely driven by processes intrinsic to the atmosphere [*Thompson et al.*, this volume]. *Feldstein* [2000] concluded, however, that while interannual variability of the NAO arises primarily from climate noise, a role for external forcings (e.g., the ocean) couldn't be entirely ruled out; for instance, about 60% of the NAO interannual variability over the last half of the 20th century is in excess of that expected if all the interannual variability was due to intraseasonal stochastic processes. *Czaja et al.* [this volume] discuss all of these issues in much more detail, and reach a similar conclusion. They argue that the ocean can be expected to modulate NAO variability on interannual and longer time scales [see also *Visbeck et al.*, this volume], and that spectral analyses of dynamical NAO indices as in Figure 12 are not optimal for detecting the impact of the ocean.

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 brief overview of these impacts. More detailed discussions can be found particularly in *Jones et al.* [this volume], *Visbeck et al.* [this volume], *Mysterud et al.* [this volume], *Drinkwater et al.* [this volume], and *Straile et al.* [this volume].

Power Spectrum   NAO Index

DJFM 1899-2002

r (1 yr) = 0.24

50   10   5   3   2

Period (years)
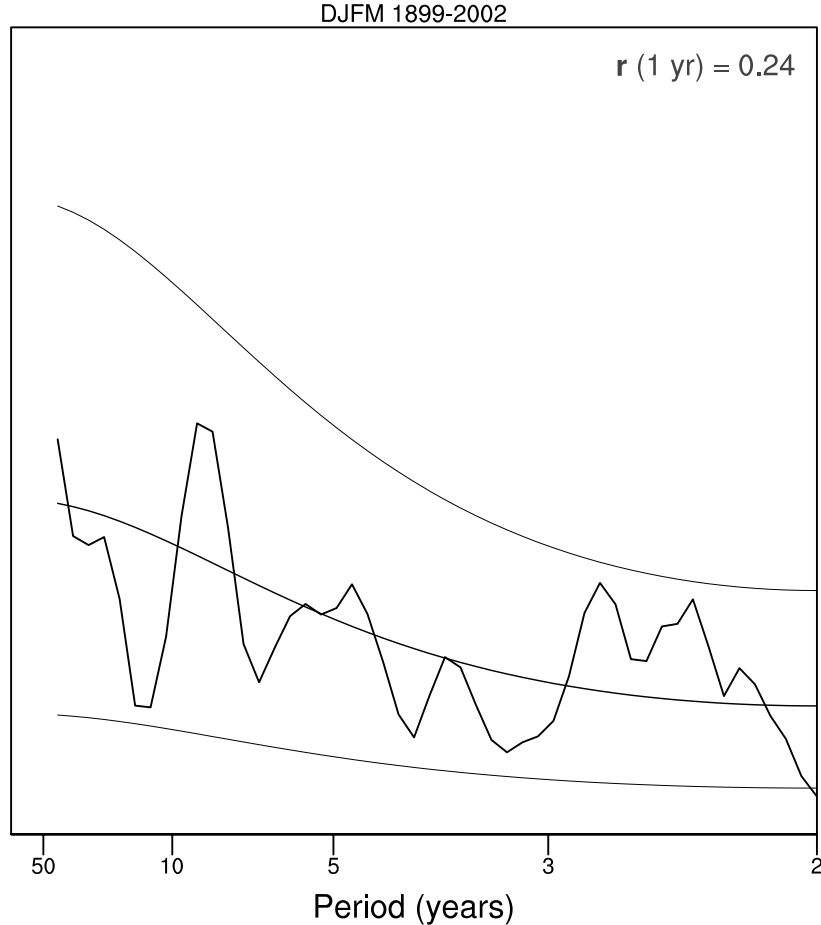
Figure 12. Power spectrum of the mean winter (December-March) NAO index over 1899-2002, defined as in the middle panel of Figure 10. Also shown is the corresponding red noise spectrum with the same lag one autocorrelation coefficient (0.24) and the 5 and 95% confidence limits.
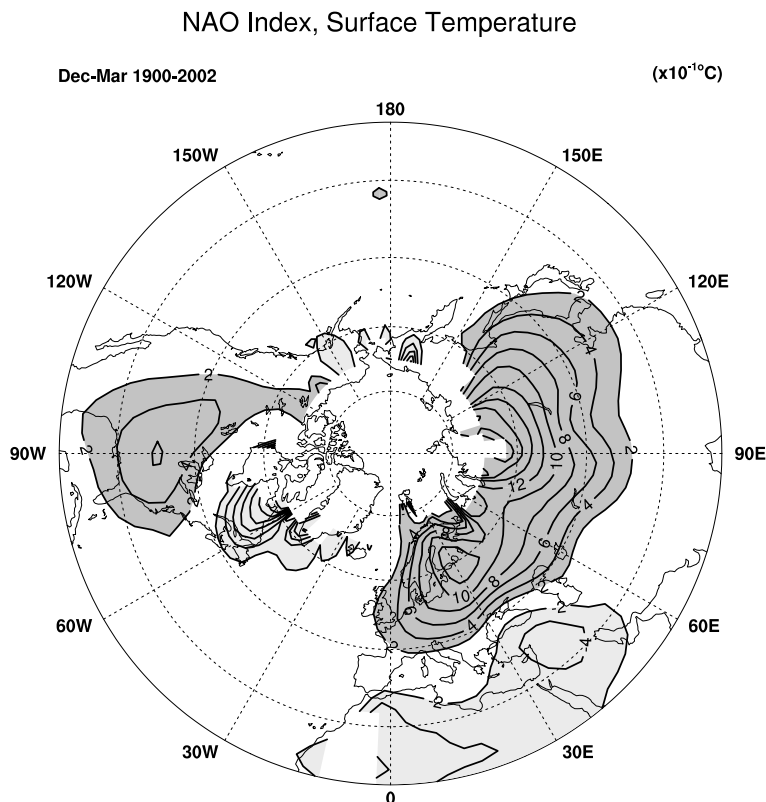
5.1. Surface Temperature

When the NAO index is positive, enhanced westerly flow across the North Atlantic during winter (Figure 7) moves relatively warm (and moist) maritime air over much of Europe and far downstream, while stronger northerly winds over Greenland and northeastern Canada carry cold air southward and decrease land temperatures and SST over the northwest Atlantic (Figure 13). Temperature variations over North Africa and the Middle East (cooling), as well as North America (warming), associated with the stronger clockwise flow around the subtropical Atlantic high-pressure center are also notable.

This pattern of temperature change is important. Because the heat storage capacity of the ocean is much greater than that of land, changes in continental surface temperatures are much larger than those over the oceans, so they tend to dominate average NH (and global) temperature variability [e.g., *Wallace et al.*, 1995; *Hurrell and Trenberth*, 1996]. Given the especially large and coherent NAO signal across the Eurasian continent from the Atlantic to the Pacific, it is not surprising that NAO variability contributes significantly to interannual and longer-term variations in NH surface temperature during winter. *Jones et al.* [this volume] show that the strength of this relationship can change over time, both locally and regionally. This aspect has implications for proxy-based reconstructions of past NAO variability [*Cook*, this volume].

Much of the warming that has contributed to the often-cited global temperature increases of recent decades has occurred during winter and spring over the northern continents [*Folland et al.*, 2001]. Since the early 1980s, winter

NAO Index, Surface Temperature

Figure 13. Changes in mean winter (December-March) land surface and sea surface temperatures (x 10^{-1}°C) corresponding to a unit deviation of the NAO index (defined as in the middle panel of Figure 10) over 1900-2002. The contour increment is 0.2°C. Temperature changes > 0.2°C are indicated by dark shading, and those < -0.2°C are indicated by light shading. Regions of insufficient data (e.g., over much of the Arctic) are not contoured, and the zero contour has been excluded.

temperatures have been 1-2°C warmer-than-average over much of North America and from Europe to Asia, while temperatures over the northern oceans have been slightly colder-than-average (Figure 14, upper panel). This pattern is strongly related to changes in the atmospheric circulation, which are reflected by lower-than-average SLP over the middle and high latitudes of the North Pacific and North Atlantic, as well as over much of the Arctic, and higher-than-average SLP over the subtropical Atlantic (Figure 14, lower panel). The Atlantic sector SLP changes clearly reflect the predominance of the positive index phase of the NAO over this period (Figure 10), while the North Pacific changes correspond to an intensification of the Aleutian low-pressure system (and an enhancement of the middle tropospheric PNA pattern) driven, at least in part, by decadal variations in ENSO [e.g., *Trenberth and Hurrell*, 1994].

Hurrell [1996] used multivariate linear regression to quantify the influence of atmospheric circulation variability associated with the NAO and ENSO on NH winter-mean surface temperatures. He showed that much of the local cooling in the northwest Atlantic and the warming across

Europe and downstream over Eurasia (Figure 14) resulted directly from decadal changes in the North Atlantic atmospheric circulation in the form of the NAO [see also *Thompson et al.*, 2000], and that the NAO (ENSO) accounted for 31% (16%) of the wintertime interannual variance of NH extratropical temperatures over the latter half of the 20th century. Moreover, changes in the atmospheric circulation associated with the NAO and ENSO accounted (linearly) for much, but not all, of the hemispheric warming through the mid-1990s [*Hurrell*, 1996]. The warming of the most recent winters, however, is beyond that that can be linearly explained by changes in the NAO or ENSO (not shown). Over 1999-2002, for instance, record warmth was recorded while generally cold conditions prevailed in the tropical Pacific and NAO-related circulation anomalies were weak.

5.2. 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,

Dec-Mar 1981-2002
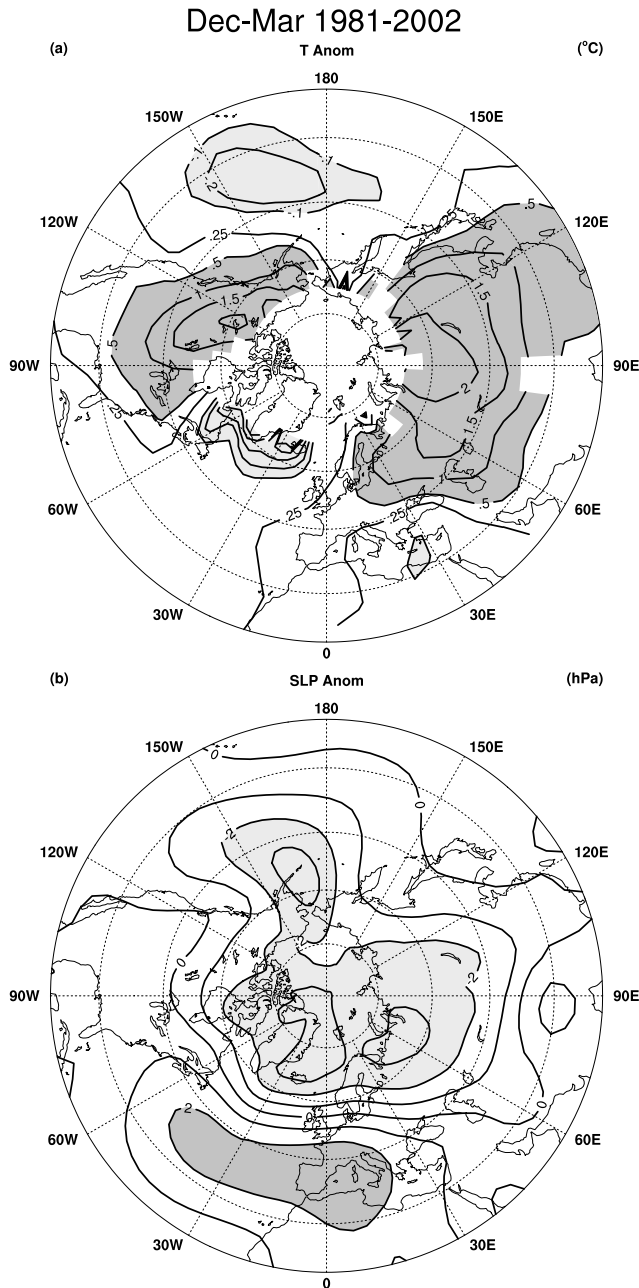
Figure 14. Twenty-two (1981-2002) winter (December-March) average (a) land surface and sea surface temperature anomalies and (b) sea level pressure anomalies expressed as departures from the 1951-1980 means. Temperature anomalies > 0.25°C are indicated by dark shading, and those < -0.10°C are indicated by light shading. The contour increment is 0.1°C for negative anomalies, and the 0.25, 0.5, 1.0, 1.5, and 2.0°C contours are plotted for positive anomalies. Regions with insufficient temperature data are not contoured. The same shading convention is used for sea level pressure, but for anomalies greater than 2 hPa in magnitude. The contour increment in (b) is 1 hPa.

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 (Figure 15). 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 (Figure 15) with enhanced activity from Newfoundland into northern Europe and a modest 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 integrates the effects of storms in the form of surface waves, so that it exhibits a marked response to long lasting shifts in the storm climate. The recent upward trend toward more positive NAO index winters has been 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 distribution of evaporation (E) and precipitation (P) [*Hurrell,* 1995a; *Dickson et al.*, 2000]. Evaporation exceeds precipitation over much of Greenland and the Canadian Arctic during high NAO index winters (Figure 16), 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*, 1995a; *Dai et al.*, 1997; *Dickson et al.*, 2000; *Visbeck et al.*, this volume].

This spatial pattern, together with the upward trend in the NAO index since the late 1960s   (Figure 10), is consistent with recent observed changes in precipitation over much of the Atlantic basin. One of the few regions of the world where glaciers have not exhibited a pronounced retreat over the past several decades is in Scandinavia [e.g., *Hagen*, 1995; *Sigurdsson and Jonsson*, 1995) where more than average amounts of precipitation have been typical of many winters since the early 1980s. In contrast, over the Alps, snow depth and duration in many recent winters have been among the lowest recorded this century, and the retreat of Alpine glaciers has been widespread [e.g., *Frank*, 1997].

Mean Storm Track (DJFM) 1958-1998

$(Z'^2)^{1/2}$ 300 hPa (gpm)
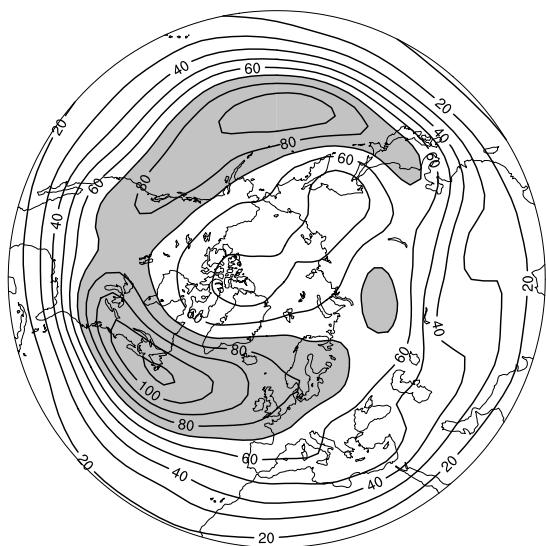
Regression onto NAO Index   (gpm)
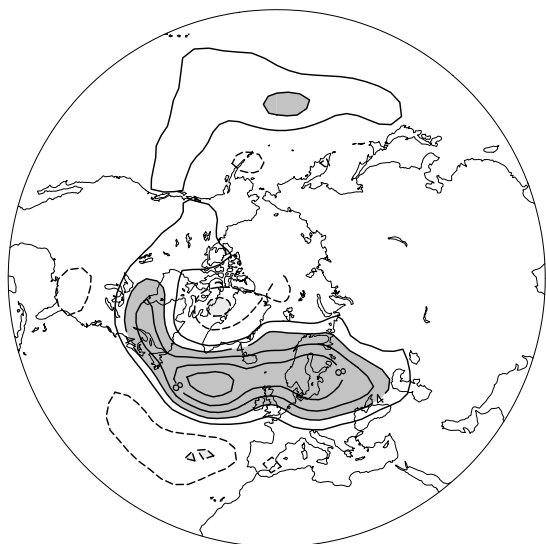
Figure 15. In the top panel, mean storm tracks for 1958-1998 winters (December-March) as revealed by the 300 hPa root mean square transient geopotential height (gpm) bandpassed to include 2-8 day period fluctuations. Values greater than 70 gpm are shaded and the contour increment is 10 gpm. In the lower panel, anomalies are expressed in terms of amplitude (gpm) by regression onto the NAO index (defined as in the middle panel of Figure 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.

Severe drought has persisted throughout parts of Spain and Portugal as well. As far eastward as Turkey, river runoff is significantly correlated with NAO variability [*Cullen and deMenocal*, 2000].

5.3. Ocean and Sea Ice

It has long been recognized that fluctuations in SST and the strength of the NAO are related, and there are clear indications that the North Atlantic Ocean varies significantly with the overlying atmosphere. *Visbeck et al.* [this volume] describe in detail the oceanic response to NAO variability.

The leading pattern of SST variability during boreal winter 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 [e.g., *Cayan*, 1992a,b; *Visbeck et al.,* this volume]. This structure suggests the SST anomalies are driven by changes in the surface wind and air-sea heat exchanges associated with NAO variations. The relationship is 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., *Battisti et al.*, 1995; *Delworth*, 1996; *Deser and Timlin*, 1997]. Over longer periods, persistent SST anomalies also appear to be related to persistent anomalous patterns of SLP (including the NAO), although a number of different mechanisms can produce SST changes on decadal and longer time scales [e.g., *Kushnir*, 1994]. Such fluctuations could primarily be the local oceanic response to atmospheric decadal variability. It is quite likely, for instance, that sustained NAO forcing results in a hemispheric SST response, in which the northern and subtropical parts of the tri-polar pattern merge [*Visbeck et al.,* this volume]. On the other hand, non-local dynamical processes in the ocean could also be contributing to the SST variations [e.g., *Visbeck et al.,* 1998; *Krahmann et al.,* 2001; *Eden and Willebrand,* 2001].

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].

The oceanic response to NAO variability is also evident in changes in the distribution and intensity of winter convective activity in the North Atlantic. The convective renewal of intermediate and deep waters in the Labrador Sea and the

High-Low NAO Index Years

E-P   December-March   (mm.day⁻¹)
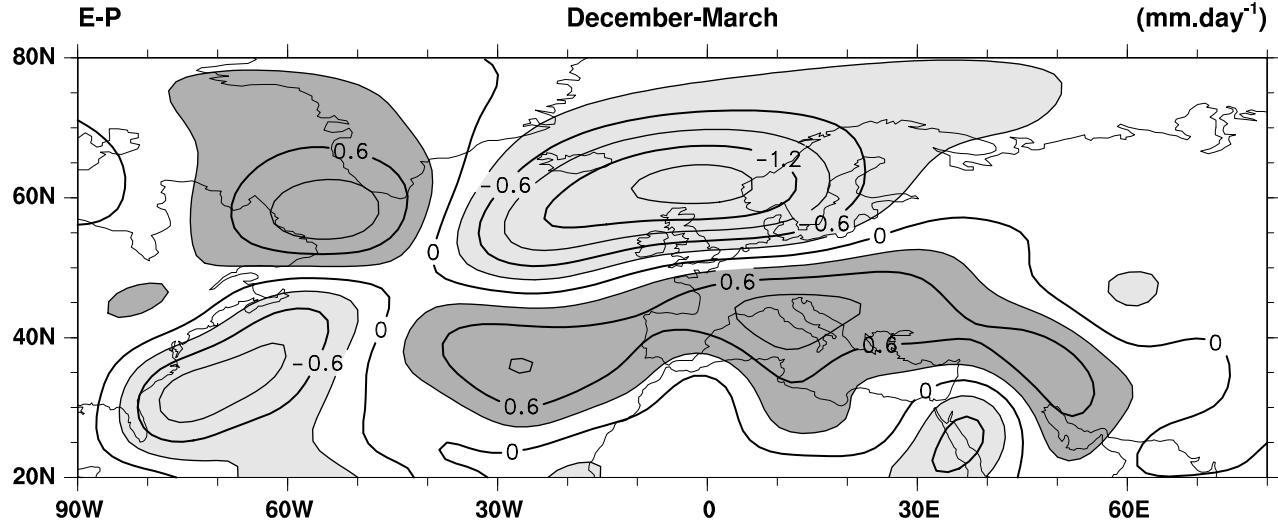
Figure 16. Difference in mean winter (December-March) evaporation (E) minus precipitation (P) between years when the NAO index exceeds one standard deviation. The NAO index is defined as in the middle panel of Figure 10, and nine winters enter into both the high index and the low index composites. The E-P field is obtained as a residual of the atmospheric moisture budget [see *Hurrell*, 1995a]. 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.

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.,* this volume; their Figure 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].

Some global warming scenarios have suggested that the next decades might show a preferred positive index phase of the NAO [*Gillett et al.*, this volume]. This would lead to increased deep water formation in the Labrador Sea region which might offset, or at least delay, the buildup of fresh water, which in many models leads to a sudden reduction of the thermohaline circulation [*Delworth and Dixon*, 2000; *Cubasch et al.,* 2001].

For this reason there has also been considerable interest in the past occurrences of low salinity anomalies that propagate around the subpolar gyre of the North Atlantic. The most famous example is the Great Salinity Anomaly (GSA) [*Dickson et al.*, 1988]. The GSA formed during the extreme negative index phase of the NAO in the late 1960s (Figure 10), when clockwise flow around anomalously high pressure over Greenland fed record amounts of freshwater from the Arctic Ocean through the Fram Strait into the Nordic Seas. From there some of the fresh water passed through the Denmark Strait into the subpolar North Atlantic Ocean gyre. There have been other similar events as well, and statistical analyses have revealed that the generation [*Belkin et al.*, 1998] and termination [*Houghton and Visbeck*, 2002] of these propagating salinity modes are closely connected to a pattern of atmospheric variability strongly resembling the NAO.

The strongest interannual variability of Arctic sea ice occurs in the North Atlantic sector. The sea ice fluctuations display 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
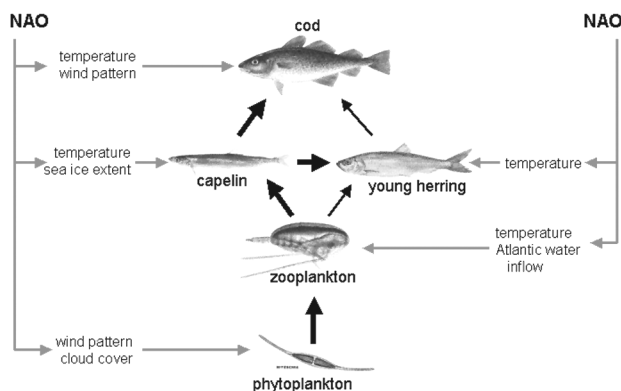
Figure 17. An example of the ecological impact of the NAO, adapted from *Stenseth et al.* [2002]. It represents a simplified food web for the Barents Sea including phytoplankton, zooplankton, capelin (*Mallotus villosus)*, herring (*Clupea harengus*), and cod (*Gadus morhua*). Positive index phases of the NAO affect the Barents Sea through increasing volume flux of warm water from the southwest, cloud cover and air temperature, all leading to increased water temperature, which influences fish growth and survival both directly and indirectly [*Ottersen and Stenseth*, 2001].

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. 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.

5.4. Ecology

Over the last couple of years interest in the ecological impacts of NAO variability has increased markedly [e.g., *Ottersen et al.*, 2001; *Walther et al.*, 2002; *Stenseth et al.*, 2002]. *Drinkwater et al.* [this volume], *Mysterud et al.* [this volume] and *Straile et al.* [this volume] show the NAO affects a broad range of marine, terrestrial and freshwater ecosystems across large areas of the NH, diverse habitats and different trophic levels. Although such effects are far-reaching, the nature of the impacts varies considerably.

Ottersen et al. [2001] attempted to systematize the ecological effects of NAO variability, and they identified three possible pathways. The first is relatively simple with few intermediary steps, such as the effect of NAO-induced temperature changes on metabolic processes such as feeding and growth (Figure 17). Since the NAO can simultaneously warm ocean temperatures in one part of the Atlantic basin and cool them in another, its impact on a single species can vary geographically. An interesting example, described by *Drinkwater et al.* [this volume], is the out-of-phase fluctuations in year-class strength of cod between the northeast and northwest Atlantic. Alternatively, more complex pathways may proceed through several physical and biological steps. One example is the intense vertical ocean mixing generated by stronger-than-average westerly winds during a positive NAO index winter. This enhanced mixing delays primary production in the spring and leads to less zooplankton, which in turn results in less food and eventually lower growth rates for fish [*Drinkwater et al.,* this volume]. A third pathway occurs when a population is repeatedly affected by a particular environmental situation before the ecological change can be perceived (biological inertia), or when the environmental parameter affecting the population is itself modulated over a number of years [physical inertia; *Heath et al.,* 1999].

Mysterud et al. [this volume] demonstrate how the NAO influences a wide range of terrestrial animals and plants, including the intriguing example of red deer on the west coast of Norway. These animals stay in low land regions during winter, while in summer they forage at higher elevations. Altitude is a key factor determining whether precipitation comes as rain or snow, and it thereby explains why positive NAO index conditions are favorable for these red deer populations. Two separate mechanisms operate. First, warm and rainy conditions in the low-elevation wintering areas decrease energetic costs of thermoregulation and movement while they increase access to forage in the field layer during winter. Second, more winter snowfall at high elevations leads to a prolonged period of access of high quality forage during summer.

Although research on the influence of the NAO on freshwater ecosystems is still in its early stages, *Straile et al.* [this volume] show that a pronounced effect on the physics, chemistry and biology of many NH lakes and rivers is apparent. To a large extent, the strong, coherent impact of the NAO on European lakes through the year is set up in winter and early spring. This time of year is critical to lakes, as both spring turnover and the onset of stratification occur. Thus, strong variations in climate driven by the NAO exert a major impact on the distribution and seasonal development of temperature and nutrients, as well as influence the time of onset, and the rate, of plankton succession. But although the NAO strongly influences a diversity of freshwater ecosystems, the actual effects differ with altitude, latitude, size, and depth of a lake.

The studies of the NAO impact on terrestrial [*Mysterud et al.*, this volume] and freshwater [*Straile et al.*, this volume] ecosystems are mainly from the eastern side of the Atlantic, and few results are available on several large groups of animals and plants. Since the study of the ecological impacts of the NAO is still relatively new and conducted by only a few scientists, it is not yet possible to determine if the NAO influences on ecology are more pronounced over Europe, or if the studies reflect more suitable European data sets and greater interest initially by European scientists. Nevertheless, interest will no doubt continue to grow leading to many new insights [*Stenseth et al.*, 2002].

5.5. Economy

Significant changes in winter rainfall and temperature driven by the NAO have the potential to affect the economy. Anecdotal evidence of this is plentiful, ranging from the economic costs of increased protection of coastal regions and the redesign of off shore oil platforms to cope with the recent NAO-driven increases in significant wave heights [e.g., *Kushnir et al.*, 1997] to altered tourism associated with changed snow conditions in the Alps and winter weather in northern Mediterranean countries (section 5.2). Some studies have also suggested the likelihood of major hurricanes land falling on the east coast of the U.S. depends on the phase of the NAO [e.g., *Elsner and Kocher*, 2000; *Elsner et al.*, 2000].

Cullen and deMenocal [2000] studied the connection between the NAO and stream flow of the Euphrates and Tigris rivers. A major issue in this region involves water supply shortages and surpluses for irrigation farming in the Middle East. Decreases in rainfall associated with the long term trend in the NAO index have had catastrophic effects on crop yields and have contributed to high level political disputes on water withdrawals from the rivers between Turkey, where most of the rain falls, and Syria, a downstream riparian neighbor.

A more recent study [J. Cherry, H. Cullen and others, personal communication] investigated the economic impact of NAO-induced changes in temperature and precipitation over Scandinavia. Norway, and its energy trade with Sweden, was examined because the Norwegian energy sector has qualities that are particularly useful for analyzing the relationship between climate and energy commodities. Norway is the leading Organization of Economic Cooperation and Development (OECD) exporter of oil and gas, and it is the second largest oil exporter in the world. Norway's climate and topography also make it ideal for hydroelectric power generation. More than 99% of electricity generation in Norway comes from hydropower, and Norway has the highest electricity consumption per capita in the world. Sweden, however, generates only 47% of its electricity from hydropower, and another 47% comes from nuclear facilities. Sweden normally adjusts its nuclear electricity production downward during the period from March to October because electricity consumption is low and water reservoirs produce plenty of hydropower. Both Norway and Sweden underwent market deregulation and increased privatization during the 1990s. This made international trade in electricity feasible in Scandinavia. In 1993, Nord Pool was established as the world's first multinational exchange for electric power trading.

In Norway the reservoir influx and level show a pronounced seasonal cycle with much month-to-month variability (Figure 18). The peak inflow in summer is associated with snowmelt in the mountain ranges. At the inflow maximum in June 1995, enough water to generate 10,000 GWh per week of hydroelectric power flowed into Norwegian reservoirs, while only enough water to generate about 5,000 GWh per week flowed into Swedish reservoirs. At the summer inflow peak in 1996, the numbers dropped to 7,000 and 4,500 GWh per week for Norway and Sweden, respectively.

This dramatic reduction in Norwegian precipitation is reflected in the strongly negative NAO index value of the December 1995-March 1996 winter, which followed a long run of wet winters (and hydropower surpluses) beginning in the late 1980s associated with the positive index phase of the NAO (Figures 10 and 16). There is a strong correlation between the winter NAO index and the amount of rainfall over western Norway (Figure 18), for instance $r \sim 0.8$ at Bergen [*Hurrell*, 1995a]. Rainfall anomalies are clearly evident in reservoir levels, which for instance were 40% below average in 1996 (Figure 18), and ultimately in hydroelectric power generation. The approximate 6-month time lag between the winter precipitation and reservoir level anomalies illustrates that much of the freshwater is stored in the winter snow pack.

The large swing in the phase of the NAO between 1995 and 1996 brought international attention to the physical connection between the NAO and the availability of water in Scandinavia for hydropower generation. Norwegian power producers sold contracts during 1995 to end-users both locally and abroad. But, during the relatively dry conditions of 1996, these contracts could not be met with Norwegian hydropower alone: power had to be purchased on short-term markets at high cost to the industry and consumers. Still embroiled in its energy crisis early in 1997, Norway imported a large amount of electricity from coal-fired power plants in Denmark, straining the commitment both countries made under the Kyoto Protocol.

Because the climates in Norway and Sweden are impacted somewhat differently by the NAO, and the electricity generation sources are distributed differently between hydropow-
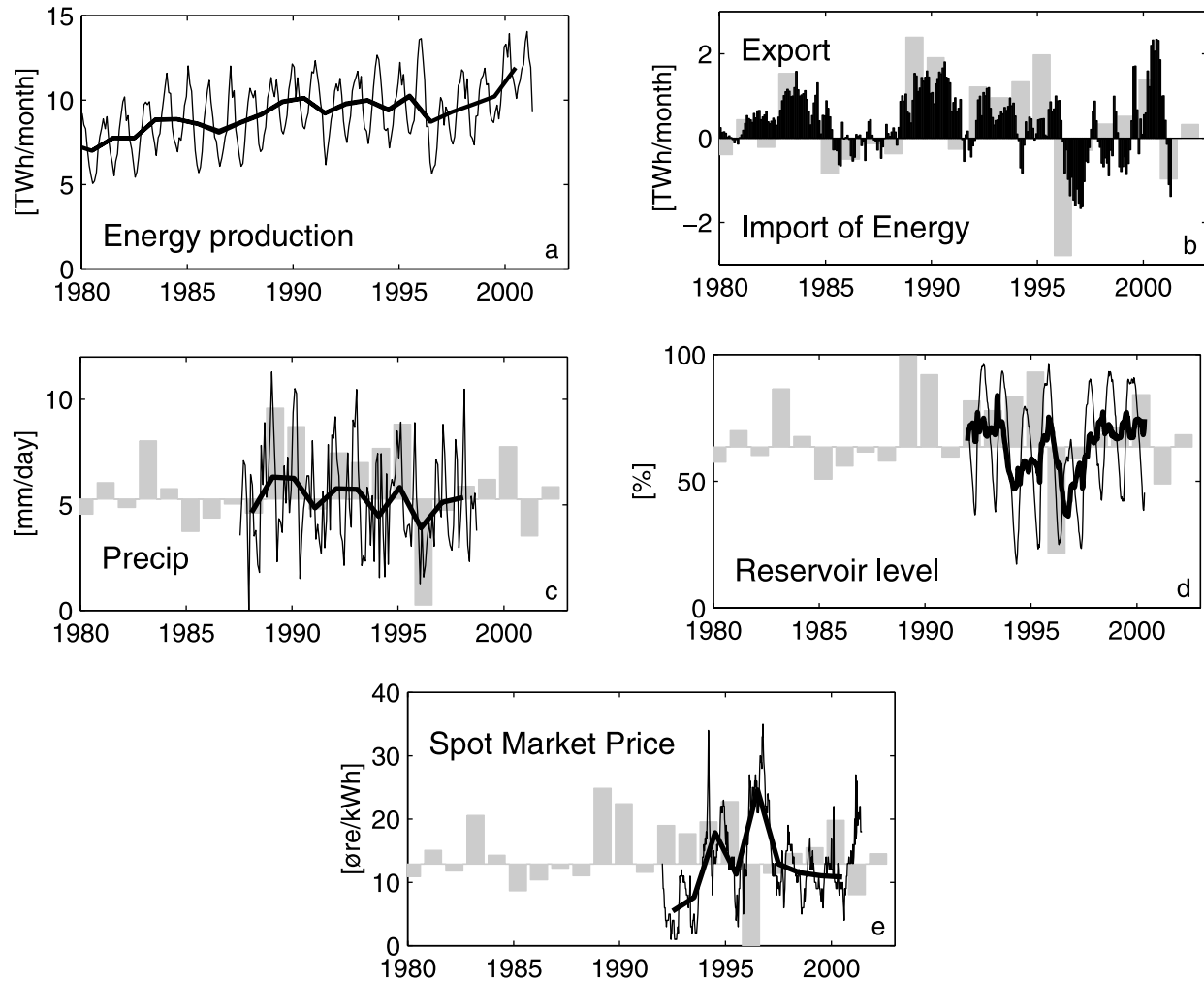
Figure 18. a) Monthly hydroelectric energy production in TWh per month for Norway (thin line). The heavy line denotes the annual average. b) Difference between hydroelectric energy production and electric energy consumption in Norway. Positive values represent times when Norway is able to export energy to Sweden and negative values indicate a shortage in hydropower production. The heavy gray bars in the background show the winter mean NAO index for this and all of the following graphs. c) Monthly time series of the first empirical orthogonal function of rainfall obtained from a gridded station based data set for all of Scandinavia (thin line). Norway has the largest spatial loading in the area where most of the hydroelectric dams are located. The heavy line denotes the annual average. d) Monthly average reservoir level for all hydroelectric dams in Norway in % (thin line). The heavy line denotes the departure from the mean annual cycle with the total time mean added back. e) Monthly price of electric energy on the Scandinavian energy spot market (thin line). The solid represents the annual average.

er, nuclear power, and fossil fuels, each country may have a natural competitive advantage under particular climate events. If the NAO is in a negative index phase and winter precipitation is relatively low, power producers in Norway can meet supply contracts by buying nuclear power from Sweden. In contrast, when the NAO is in its positive index phase and hydroelectric power is plentiful in Norway, Sweden may find it cheaper to buy hydropower from Norway than to produce that power itself. The existence of electricity trade means that Norway can maintain its heavy reliance on hydropower instead of burning fossil fuels. Figure 18 shows how, in recent years, the NAO has influenced both the trade amount and price of energy on the spot market.

6. MECHANISMS

Although the overwhelming indication is that the NAO is a mode of variability internal to the atmosphere, there is

some evidence that external factors such as volcanic aerosols, anthropogenic influences on the atmospheric composition, and variations in solar activity can influence its phase and amplitude. Moreover, it has been argued that interactions between the atmosphere and the underlying surface, or between the troposphere and stratosphere, can lend a "low-frequency" component to the NAO variability, such that limited prediction is plausible. At present there is no consensus on the relative roles such processes play in NAO variability, especially where long (interdecadal) times scales are concerned. Considering the significant impact the NAO exerts on the climate of the NH, understanding the mechanisms that control and affect the NAO is therefore crucial to the current debate on climate variability and change. *Thompson et al.* [this volume], *Czaja et al.* [this volume], and *Gillett et al.* [this volume] provide more detailed discussions of external physical processes thought to affect the NAO in the context of thorough overviews of the dynamical factors that give rise to the horizontal and vertical structure of NAO variability, as well as its amplitude and time scales. Here we provide a quick look at these mechanisms and the debate surrounding their importance.

6.1. Atmospheric Processes

Atmospheric general circulation models (AGCMs) provide strong evidence that the basic structure of the NAO arises from the internal, nonlinear dynamics of the atmosphere. The observed spatial pattern and amplitude of NAO anomalies are well simulated in AGCMs forced with climatological annual cycles (no interannual variations) of all forcings "external" to the atmosphere, like insolation, SST, sea ice, snow cover, and land surface moisture, as well as fixed atmospheric trace-gas composition [e.g., *Barnett*, 1985; *James and James*, 1989; *Kitoh et al.*, 1996; *Osborn et al.*, 1999]. The results from one such integration are illustrated in Figure 19. *Thompson et al.* [this volume] discuss in depth the governing atmospheric dynamical mechanisms, of which interactions between the time-mean flow and the departures from that flow (the so-called transient eddies) are central and give rise to a fundamental time scale for NAO fluctuations of about 10 days [*Feldstein*, 2000]. Such intrinsic atmospheric variability exhibits little temporal coherence (Figure 19), mostly consistent with the time scales of observed NAO variability (Figures 10 and 12) and the climate noise paradigm discussed earlier (section 4.2).

A possible exception to this reference is the enhanced NAO variability over the latter half of the 20th century [*Feldstein*, 2000], including the apparent upward trend in the boreal winter NAO index (Figure 19). *Thompson et al.*

[2000] concluded that the linear component of the observed upward trend of the index is statistically significant relative to the degree of internal variability that it exhibits, and *Gillett et al.* [2001] and *Feldstein* [2002] showed that the upward trend is statistically significant compared to appropriate red noise models [*Trenberth and Hurrell*, 1999; although see *Wunsch,* 1999]. *Osborn et al.* [1999] showed, furthermore, that the observed trend in the winter NAO index is outside the 95% range of internal variability generated in a 1,400 year control run with a coupled ocean-atmosphere climate model, and *Gillett et al.* [this volume] reach the same conclusion based on their examination of multi-century control runs from seven different coupled climate models. This indicates that either the recent climate change is due in part to external forcing, or all of the models are deficient in their ability to simulate North Atlantic interdecadal variability (although the simulated variability was similar to that observed in the instrumental record prior to 1950). Comparisons to NAO indices reconstructed from proxy data have also concluded the recent behavior is unusual, although perhaps not unprecedented [*Jones et al.* 2001; *Cook*, this volume]; however, such extended proxy records are liable to considerable uncertainties [e.g., *Schmutz et al.,* 2000; *Cook*, this volume].

A possible source of the trend in the winter NAO index could be external processes that affect the strength of the atmospheric circulation in the lower stratosphere on long time scales, such as increases in GHG concentrations [*Gillett et al.*, this volume]. In contrast to the mean flow (Figures 1-3), the NAO has a pronounced "equivalent barotropic" structure (i.e., it does not display a westward tilt with elevation), and its anomalies increase in amplitude with height in rough proportion to the strength of the mean zonal wind [*Thompson et al.*, this volume]. In the lower stratosphere, the leading pattern of geopotential height variability is characterized by a much more annular (zonally symmetric) structure than in the troposphere. When heights over the polar region are lower than normal, heights at nearly all longitudes in middle latitudes are higher than normal, and *vice versa*. In the former phase, the stratospheric westerly winds that encircle the pole are enhanced and the polar vortex is "strong" and anomalously cold. Simultaneously at the surface, the NAO tends to be in its positive index phase [e.g., *Baldwin*, 1994; *Perlwitz and Graf*, 1995; *Kitoh et al.*, 1996; *Kodera et al.*, 1996; *Baldwin and Dunkerton*, 1999].

The trend in the NAO phase and strength during the last several decades has been associated with a stratospheric trend toward much stronger westerly winds encircling the pole and anomalously cold polar temperatures [e.g., *Randel and Wu*, 1999; *Thompson et al.*, 2000]. There is a considerable body of research and observational evidence to support

the notion that variability in the troposphere can drive variability in the stratosphere, but it now appears that some stratospheric control of the troposphere may also occur [e.g., *Baldwin and Dunkerton*, 2001]. *Thompson et al.* [this volume] review the evidence for such "downward control" and evaluate the possible mechanisms, which likely involve the effect of the stratospheric flow on the refraction of planetary waves dispersing upwards from the troposphere [e.g., *Hartmann et al.*, 2000; *Shindell et al.*, 2001; *Ambaum and Hoskins*, 2002], although more direct momentum forcing could also be important [e.g., *Haynes et al.*, 1991; *Black*, 2002].

The atmospheric response to strong tropical volcanic eruptions provides some evidence for a stratospheric influence on the surface climate [*Gillett et al.*, this volume]. Volcanic aerosols act to enhance north-south temperature gradients in the lower stratosphere by absorbing solar radiation in lower latitudes, which produces warming. In the troposphere, the aerosols exert only a very small direct influence [*Hartmann et al.*, 2000]. Yet, the observed response following eruptions does not include only lower geopotential heights over the pole with stronger stratospheric westerlies, but also a strong, positive NAO-like signal in the tropospheric circulation [e.g., *Robock and Mao*, 1992; *Kodera*, 1994; *Graf et al.*, 1994; *Kelley et al.*, 1996].

Reductions in stratospheric ozone and increases in GHG concentrations also appear to enhance the meridional temperature gradient in the lower stratosphere, via radiative cooling of the wintertime polar regions. This change implies a stronger polar vortex. It is possible, therefore, that the upward trend in the boreal winter NAO index in recent decades (Figure 10) is associated with trends in either or both of these trace-gases quantities. In particular, a decline in the amount of ozone poleward of $40°N$ has been observed during the last two decades [*Randel and Wu*, 1999].

Shindell et al. [1999] subjected a coupled ocean-atmosphere model to realistic increases in GHG concentrations and found a trend toward the positive index phase of the NAO in the surface circulation. In similar experiments with different coupled models, *Ulbrich and Christoph* [1999] and *Fyfe et al.* [1999] also found an increasing trend in the NAO index, although *Osborn et al.* [1999] and *Gillett et al.* [2000] did not. This led *Cubasch et al.* [2001] to conclude that there is not yet a consistent picture of the NAO response to increasing concentrations of greenhouse gases. *Gillett et al.* [this volume], however, examine 12 coupled ocean-atmosphere models and find 9 show an increase in the boreal winter NAO index in response to increasing GHG levels, although the results are somewhat sensitive to the definition of the NAO index used (section 3). This led them to conclude that increasing GHG concentrations have contributed to a strengthening of the North Atlantic surface pressure gradient [see also *Osborn*, 2002]. Forcing from stratospheric ozone depletion has generally been found to have a smaller effect than GHG changes on the NAO [e.g., *Graf et al.*, 1998; *Shindell et al.*, 2001], although *Volodin and Galin* [1999] claim a more significant influence. *Gillett et al.* [this volume] synthesize the diverse and growing body of literature dealing with how the NAO might change in response to several anthropogenic forcings.

6.2. The Ocean's Influence on the NAO

In the extratropics, the atmospheric circulation is the dominant driver of upper ocean thermal anomalies [section 5.3; *Visbeck et al.*, this volume]. A long-standing issue, however, has been the extent to which the forced, anomalous extratropical SST field feeds back to affect the atmosphere [*Kushnir et al.*, 2002]. Most evidence suggests this effect is quite small compared to internal atmospheric variability [e.g., *Seager et al.*, 2000]. Nevertheless, the interaction between the ocean and atmosphere could be important for understanding the details of the observed amplitude of the NAO and its longer-term temporal evolution [*Czaja et al.*, this volume], as well as the prospects for meaningful predictability [*Rodwell et al.*, this volume].

The argument for an oceanic influence on the NAO goes as follows: while intrinsic atmospheric variability exhibits temporal incoherence, the ocean tends to respond to it with marked persistence or even oscillatory behavior. The time scales imposed by the heat capacity of the upper ocean, for example, leads to low frequency variability of both SST and lower tropospheric air temperature [*Frankignoul and Hasselmann*, 1977; *Manabe and Stouffer*, 1996; *Barsugli and Battisti*, 1998]. Other studies suggest that basin-wide, spatially-coherent atmospheric modes such as the NAO interact with the mean oceanic advection in the North Atlantic to preferentially select quasi-oscillatory SST anomalies with long time scales [*Saravanan and McWilliams*, 1998] and even excite selected dynamical modes of oceanic variability that act to redden the SST spectrum [*Griffies and Tziperman*, 1995; *Frankignoul et al.*, 1997; *Capotondi and Holland*, 1997; *Saravanan and McWilliams*, 1997; 1998; *Saravanan et al.*, 2000]. These theoretical studies are supported by observations that winter SST anomalies, born in the western subtropical gyre, spread eastward along the path of the Gulf Stream and North Atlantic Current with a transit time of roughly a decade [*Sutton and Allen*, 1997; *Krahmann et al.*, 2001]. *Czaja et al.* [this volume] offer a detailed examination of all these possibilities of ocean-atmosphere interaction.
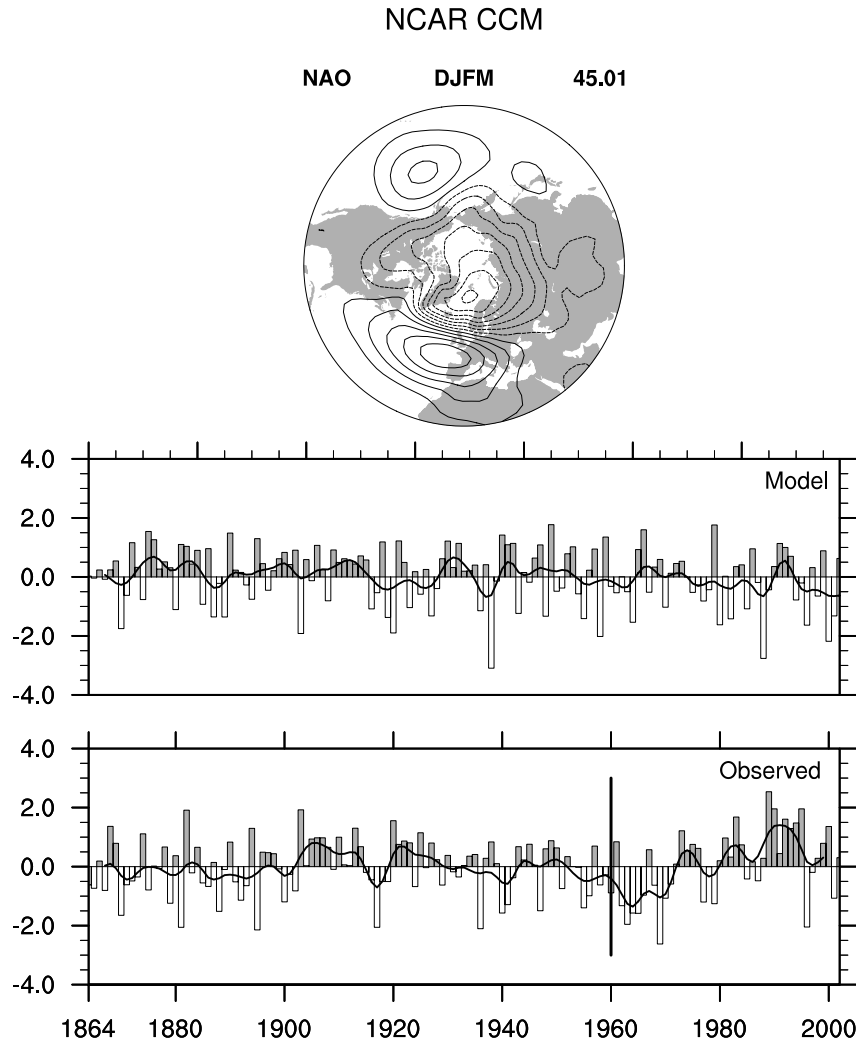
Figure 19. Leading empirical orthogonal function (EOF 1) of the winter (December-March) mean sea level pressure anomalies over (top) the North Atlantic sector (20º-70ºN, 90ºW-40ºE), and the percentage of the total variance it explains in a 200 year integration with the NCAR Community Climate Model (CCM) with climatological annual cycles of all forcings external to the atmosphere. The pattern is displayed in terms of amplitude (hPa), obtained by regressing the hemispheric sea level pressure anomalies upon the leading principal component time series. A subset of 139 years of the latter is plotted in the middle panel for comparison to the station-based NAO index (lower panel) of equivalent length (upper panel of Figure 10). Both the model and observed NAO indices have been normalized. In the top panel, the contour increment is 0.5 hPa, and the zero contour has been excluded.

A key question in this debate is the sensitivity of the middle latitude atmosphere, away from the surface, to changes in SST (and other surface boundary conditions including sea-ice and land surface snow cover). This issue has been addressed in numerous studies, many of them based on AGCM experiments with prescribed SST anomalies. *Palmer and Sun* [1985] and *Peng et al.* [1995], for instance, showed that a stationary, warm SST anomaly south of Newfoundland, in the Gulf Stream Extension region, forces a high-pressure response over the North Atlantic, but in spatial quadrature with the NAO pattern. When a more realistic reproduction of the entire basin SST response to NAO variability [*Visbeck et al.*, this volume] is prescribed, a realistic atmospheric NAO pattern emerges as a response, but whether the forcing comes from the tropical or extratropical part of the ocean has not been unequivocally resolved [*Venzke et al.*, 1999; *Sutton et al.*, 2001; *Peng et al.*, 2002]. *Robertson et al.* [2000] reported that changing the SST distribution in the North Atlantic affects the frequency of occurrence of different regional low-frequency atmospher-

ic modes and substantially increases the interannual variability of the NAO simulated by their AGCM. *Rodwell et al.* (1999) show that by forcing their AGCM with globally-varying observed SST and sea ice distributions, the phase (though not the full amplitude) of the long-term variability in the observed, wintertime NAO index over last half century can be captured, including about 50% of the observed strong upward trend over the past 30 years [see also *Mehta et al.*, 2000; *Hoerling et al.*, 2001].

The weakness of AGCM responses to SST anomalies and the occasional confusing and inconsistent results [e.g., *Kushnir and Held,* 1996] created a debate as to the importance of oceanic forcing for climate anomalies in general and the NAO in particular [*Kushnir et al.*, 2002]. It is possible that AGCM experiments with prescribed SST do not correctly represent the processes in nature, where atmospheric fluctuations cause the SST variability, and it is the "back interaction" or feedback of the SST anomalies that is sought. *Barsugli and Battisti* [1998] argued that when the ocean responds to the atmosphere, the thermal damping on the latter is reduced, thereby contributing to stronger and more persistent anomalies in both media. *Kushnir et al.* [2002] proposed a paradigm, based on North Pacific SST studies [*Peng et al.,* 1997; *Peng and Whitaker*, 1999], in which the changes in SST due to large scale circulation anomalies (such as the NAO) modify the temperature gradient at the surface and, thus, the associated baroclinic transient activity, which in turn feeds back on the large scale flow anomalies [e.g., *Hurrell* 1995b]. This mechanism may act together with reduced thermal damping to explain why the NAO is more persistent during winter (and well into March) than during the rest of the year, and why there is a high correlation between late spring SST anomalies and the NAO state in the ensuing fall and early winter [*Czaja and Frankignoul*, 2002; *Rodwell*, this volume].

In their review, *Czaja et al.* [this volume] explore the implication of the reduced thermal damping paradigm, as well as evidence from coupled general circulation model experiments that the climate system exhibits quasi-oscillatory behavior due to the long-term (multi-year) response of the ocean circulation to atmospheric forcing [e.g., *Latif and Barnett*, 1996; *Grötzner et al.*, 1998; *Timmermann et al.*, 1998]. These latter studies argue that changes in the oceanic heat transport (perhaps driven by NAO variability) can lead to a negative feedback on the atmosphere and thus reverse the phase of the atmospheric forcing phenomenon. Such behavior is obviously conditioned on the ability of the atmosphere to respond to upper ocean heat content anomalies (see discussion above). The possibility that this delayed circulation response will lead to unstable, oscillatory modes of the climate system has also been explored in simplified models [e.g., *Jin* 1997; *Goodman and Marshall*, 1999; *Weng and Neelin*, 1998].

Adding to the complexity of ocean-atmosphere interaction is the possibility of remote forcing of the NAO from the tropical oceans. Several recent studies have concluded that NAO variability is closely tied to SST variations over the tropical North and South Atlantic [*Xie and Tanimoto*, 1998; *Rajagopalan et al.*, 1998; *Venzke et al.*, 1999; *Robertson et al.*, 2000; *Sutton et al.*, 2001]. SST variations in the tropical Atlantic have large spatial dimension and occur on a wide range of time scales, from interannual to decadal. They involve changes in the meridional SST gradient across the equator, which affect the strength and location of tropical Atlantic rainfall and thus possibly influence the North Atlantic middle latitude circulation. On the other hand, *Hoerling et al.* [2001] used carefully designed AGCM experiments to argue that the multi-year trend in North Atlantic circulation toward the positive index phase of the NAO since 1950 has been driven by a commensurate warming of the tropical Indian Ocean and western Pacific surface waters. Although *Hoerling et al.* [2001] did not directly assess the role of the extratropical oceans in their paper, the fact that most of the low frequency behavior of the observed NAO index since 1950 is recoverable from tropical SST forcing alone suggests a more passive role for extratropical North Atlantic SSTs [see also *Peterson et al.*, 2002]. *Sutton and Hodson* [2002] also find evidence of tropical Indian Ocean forcing of the NAO on long time scales, but in contrast to *Hoerling et al.* [2001], they conclude this effect is secondary to forcing from the North Atlantic itself. Clearly, the importance of tropical versus extratropical ocean-atmosphere interaction has not yet been fully determined.

Similarly, the impact of ENSO on the North Atlantic climate, and the NAO in particular, remains open to debate, although most evidence suggests the effects are small but non-trivial. *Rogers* [1984] concluded from an analysis of historical SLP data that simultaneous occurrences of particular modes of ENSO and the NAO "seem to occur by chance", and this is consistent with the fact that there is no significant correlation between indices of the NAO and ENSO on interannual and longer time scales. *Pozo-Vázquez et al.* [2001], however, composited extreme ENSO events and found for cold conditions in the tropical Pacific a statistically significant boreal winter SLP anomaly pattern resembling the positive index phase of the NAO. *Cassou and Terray* [2001b] also argued for an influence of La Niña on the North Atlantic atmosphere, although they found an atmospheric response that more closely resembles the ridge regime of Figure 9. *Sutton and Hodson* [2002] point out, importantly, that the influence of ENSO most likely depends on the state of the North Atlantic itself [see also *Venzke et al.,* 1999; *Mathieu et al.*, 2002]. Moreover, the possibility of an *indirect* link between ENSO and the NAO is suggested by numerous stud-

ies that illustrate a direct impact of ENSO on tropical North Atlantic SSTs [e.g., *Enfield and Mayer*, 1997; *Klein et al.*, 1999; *Saravanan and Chang*, 2000; *Chiang et al.*, 2000; 2002]. If this were a strong effect, however, it would be detected by statistical analyses.

6.3. Sea Ice and Land Snow Cover

The role of sea ice and land snow cover in affecting atmospheric variability has received very little attention, especially relative to the role of ocean anomalies. Here, too, the issue is whether changes in surface properties due to the NAO are able to modify its phase and amplitude in turn. As discussed in section 5.3, changes in sea-ice cover in both the Labrador and Greenland Seas as well as over the Arctic are well correlated with NAO variations [*Deser et al.*, 2000]. Since changes in ice cover produce large changes in sensible and latent heat fluxes, it is reasonable to ask if there is a subsequent feedback onto the atmospheric circulation anomalies. *Deser et al.* [2000] suggest from observations that a local response of the atmospheric circulation to reduced sea ice cover east of Greenland in recent years is apparent. However, more recent AGCM experiments, with imposed ice cover anomalies consistent with the observed trend of diminishing (increasing) ice concentration during winter east (west) of Greenland, suggest a weak, negative feedback onto the NAO [C. Deser, personal communication].

Watanabe and Nitta [1999] suggested that land processes are responsible for decadal changes in the NAO. They find that the change toward a more positive wintertime NAO index in 1989 was accompanied by large changes in snow cover over Eurasia and North America. Moreover, the relationship between snow cover and the NAO was even more coherent when the preceding fall snow cover was analyzed, suggesting that the atmosphere may have been forced by surface conditions over the upstream land mass. *Watanabe and Nitta* [1998] reproduce a considerable part of the atmospheric circulation changes by prescribing the observed snow cover anomalies in an AGCM. The debate on the importance of NAO interactions with the earth land and ocean surface is thus far from over.

7. CONCLUSIONS AND CHALLENGES

The NAO is a large-scale atmospheric phenomenon primarily arising from stochastic interactions between atmospheric storms and both the climatological stationary eddies and the time mean jet stream. 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 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 influence the atmospheric circulation over the North Atlantic and, in particular, the NAO. Tropical convection 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.

Interactions with the lower stratosphere are a second possibility. This mechanism is of interest because it might also explain how changes in atmospheric composition influence the NAO. For example, changes in ozone, 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. Such interactions have also been used to rationalize the recent trend in the winter NAO index in terms of global warming.

A third possibility is that the state of the NAO is influenced by variations in heat exchange between the atmosphere and the ocean, sea-ice and/or land systems. A significant amount of numerical experimentation has been done to test the influence of tropical and extratropical SST anomalies on the NAO, and these experiments are now beginning to lead to more conclusive and coherent results.

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.

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