NORTH ATLANTIC OSCILLATION – CONCEPTS AND STUDIES

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Abstract. This paper aims to provide a comprehensive review of previous studies and concepts concerning the North Atlantic Oscillation. The North Atlantic Oscillation (NAO) and its recent homologue, the Arctic Oscillation/Northern Hemisphere annular mode (AO/NAM), are the most prominent modes of variability in the Northern Hemisphere winter climate. The NAO teleconnection is characterised by a meridional displacement of atmospheric mass over the North Atlantic area. Its state is usually expressed by the standardised air pressure difference between the Azores High and the Iceland Low. This NAO index is a measure of the strength of the westerly flow (positive with strong westerlies, and vice versa). Together with the El Niño/Southern Oscillation (ENSO) phenomenon, the NAO is a major source of seasonal to interdecadal variability in the global atmosphere. On interannual and shorter time scales, the NAO dynamics can be explained as a purely internal mode of variability of the atmospheric circulation. Interdecadal variability may be influenced, however, by ocean and sea-ice processes.

Keywords: atmospheric circulation, climate, flow, North Atlantic Oscillation

Abbreviations: AMO – Atlantic Multidecadal Oscillation; AO/NAM – Arctic Oscillation (synonym to Northern Hemisphere annular mode, NAM); AOGCM – Atmosphere-Ocean General Circulation Model; AOI – Arctic Oscillation Index; CCA – Canonical Correlation Analysis; ECMWF – European Centre for Medium-Range Weather Forecasts; ENSO – El Niño/Southern Oscillation; EOF – Empirical Orthogonal Function; EU – Eurasian Pattern; GA – Greenland Above; GB – Greenland Below; GC – Gyre Circulation; GCM – Global Circulation Model; GHG – Greenhouse Gas; GIN Seas – Greenland-Iceland-Norwegian Seas; GR – Gridded NAOI by Luterbacher et al. (1999; 2001a); GSA – Great Salinity Anomaly; HU – NAO by Hurrell (1995a); ITCZ – Intertropical Convergence Zone; JO – NAOI by Jones et al. (1997); LO – Lorenz index (Lorenz 1951); MFT – Multiresolution Fourier Transform; NAC – North Atlantic Current; NAO – North Atlantic Oscillation; NAOI – North Atlantic Oscillation Index; NCEP/NCAR – National Center for Environmental Prediction/National Center for Atmospheric Research; NG – Temperature Norway Greenland (Wallace 2000); NH – Northern Hemisphere; PC – Subpolar SLP principal component; PCA – Principal Component Analysis; PNA – Pacific North American Pattern; PNJ – Polar Night Jet; PJO – Polar-night Jet Oscillation; QBO – Quasi Biennial Oscillation or Tropical Biennial Oscillation; RO – normalised NAOI by Rogers (1984); SH – Southern Hemisphere; SIC – Sea Ice Concentration; SLP – Sea Level Pressure; SST – Sea Surface Temperature; SVD – Singular Value Decomposition; TAV – Tropical Atlantic Variability; TDS – Transpolar Drift Stream; THC – Thermohaline Circulation; TW – AOI according to Thompson and Wallace (1998); WA – Western Atlantic; WB – Walker and Bliss NAOI (1932).

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1. Introduction

Everyone living in Europe knows that different winters have different characteristics. For example, the winters between 1975 and 1995 were generally mild and there was little snow in Central Europe, whereas in the 1960s winters were often cold and dry like 1963 and sometimes large amounts of snow were observed (Wanner et al., 1997; Uppenbrink, 1999). Meteorologists would say that air temperatures and snowfall in both cases were related to the atmospheric circulation over the North Atlantic–European sector. A long lasting weather situation can at most characterise a whole winter, but what makes mild or severe winters tend to occur in succession? Some weather situations might occur over some years more frequently than others might, and then the opposite could be the case in other periods. This question is not only important for understanding climate but also raises the hope of being able to make seasonal forecasts.

One such mode that comes out of many climate statistics is often called the North Atlantic Oscillation (NAO; we will go into more detail about the concurrent modes and the ongoing debate in the next Section). The NAO describes a large-scale meridional vacillation in atmospheric mass between the North Atlantic regions of the subtropical anticyclone near the Azores and the subpolar low pressure system near Iceland. It is a major source of seasonal to interdecadal variability in the worldwide atmospheric circulation (Hurrell, 1995a) and represents the most important "teleconnection" (some authors prefer the term "anomaly pattern" in this context; Wallace and Gutzler, 1981; Kushnir and Wallace, 1989) of the North Atlantic-European area (Hurrell and van Loon, 1997; Kapala et al., 1998), where it is most pronounced in winter. The measure for the state of the NAO, the North Atlantic Oscillation Index (NAOI) is widely used as a general indicator for the strength of the westerlies over the eastern North Atlantic and western Europe and most importantly for winter climate in Europe (Hurrell and van Loon, 1997; Wanner et al., 1997; WMO, 1998). In fact, the NAOI is highly correlated with a large variety of atmosphere-related environmental variables, mainly during the winter season (see Dickson et al., 2000 and Souriau and Yiou, 2001, for nice overviews).

The NAO is a descriptive summary, and consequently its precise definition depends on the statistical approach used. Unfortunately, a uniquely defined NAOI does not exist, nor is there a unique spatial pattern. However, the NAO is not only the result of statistical considerations. What is behind the statistics, what is behind the complex NAO dynamics? Is it simply the dynamical expression of the land-sea distribution and orography of the Northern Hemisphere (NH) that inevitably has to come out from any circulation statistics? Or is it the expression of time-dependent processes, or both, a result of spatio-temporal considerations? Is the NAOI varying randomly or is it controlled on certain time scales? Does the NAO respond to forcings applied by volcanic eruptions, changes in the solar activity, greenhouse gases, or just by anomalous ocean temperatures? There is no single conceptual framework to what exactly controls the NAO. There are alternative views; a situation which characterises the current debate. A pattern that is closely related to the NAO, the Arctic Oscillation (AO) or Northern Hemisphere annular mode (NAM; Thompson and Wallace, 1998; 2001; Wallace, 2000) has recently received much attention. It is defined as the leading EOF of the monthly sea level pressure (SLP) fields north of 20° N weighted by area. The AO has large similarities to the NAO with respect to its spatial pattern and their index series are highly correlated (more details will be given in Chapter 6). NAO and AO describe the same phenomenon, but lead to different dynamical interpretations (Wallace, 2000). In a recent paper, Delworth and Dixon (2000) denote the NAO as the regional representation of the largest changes in midlatitudes associated with the AO, occurring in the Atlantic sector. In this contribution, we would like to present both views. However, since there is the danger to be judged from the use of an acronym, we want to make clear that we use the acronym NAO whenever the phenomenon is referred to (for historical reasons) and make the distinction between NAO and AO/NAM when discussing the dynamical concepts.

The NAO has received a lot of attention in the climate research community especially in recent years, and even more so with the proposition of the AO/NAM. The NAO-AO/NAM issue rivals the El Niño/Southern Oscillation in terms of its significance for understanding global climate variability (Wallace, 2000). Researchers from other fields have started to be interested in this phenomenon, and it appears more and more in the media. Despite the lack of a unifying concept, many detailed studies have been published and some new dynamical ideas have been developed in the last years (for an overview see Greatbach, 2000).

This paper aims to inform a broader scientific community and even interested non-specialists about the complex NAO-AO/NAM phenomenon. Section 2 aims to give an overview about the concept and the dynamics of the NAO or related modes. The history of its discovery and research is described in Section 3. This part of the article is designed to be understandable also for the above mentioned non-specialists. The North Atlantic Oscillation dynamics, including their spatial representation, are described in the Sections 4 and 5 by including not only observational but also theoretical and modeling studies. Section 6 presents the statistical analysis of mainly two time series of NAO indices. This part also comprises a diagnostic study of the state of the NAO since AD 1659, including a spectral analysis. Section 7 deals with the simulated future trends of the North Atlantic and Arctic Oscillations (NAO and AO/NAM) in climate models and the projected changes for the future. Conclusions from the point of view of the researcher and consequences for our understanding of climate are given in chapter 8.

2. North Atlantic Oscillation: The Phenomenon

Circulation modes such as the NAO are found by decomposing the spatio-temporal variability of atmospheric variables into spatial patterns on one hand, and time series that describe the strength of the patterns on the other. The decomposition can be based entirely on statistical methods. This is the case of the AO. It can also be based on a priori knowledge, such as the predominance of the Azores High and the Icelandic Low for weather and climate in Europe. This is the case for the NAO. It should be noted that these are not the only centres of action relevant for European weather and climate. Other modes like the Eastern Atlantic and the Eurasian Pattern (EU) are also important (see e.g., Barnston and Livezey, 1987; Cayan 1992b and Luterbacher et al., 1999).

Before entering the discussion about the concepts and the dynamical background, we would like to start the paper with a short description of the spatial manifestation of the NAO pattern in the SLP distribution and to look at the regional impact, i.e., the spatial pattern of the relation of other relevant climate variables (temperature, precipitation) to the NAO.

The SLP distribution over the North Atlantic for the positive NAO mode (NAO+) has a well developed Icelandic Low and Azores High, associated with stronger westerlies over the eastern North Atlantic and the European continent. Figure 1a shows an example of a month (January 1990) with a strong pressure gradient between Iceland and the Azores, thus a positive NAOI. In the negative NAO mode (NAO−), an example is shown in Figure 1b (February 1972), the Icelandic Low and the Azores High are rather weak, thus giving rise to reduced westerlies over the eastern North Atlantic. Note, however, that in the negative NAO mode, the pressure distribution is not necessarily reversed. There is still an Icelandic Low and an Azores High, but weaker than normal. Complete reversals with higher pressure over Iceland region than over the Azores region in a monthly average (extremely negative NAOI) occur very rarely but are particularly important. They share a remarkable part of the interdecadal variability of the surface air temperature in the North Atlantic sector (Moses et al., 1987). An example of a reversal in SLP is given in Figure 1c (January 1963) which was associated with a strong easterly flow over the eastern North Atlantic. Other winter months with typical reversals were January 1881 and January 1918.

The positive and negative phases of the NAO mode are accompanied by different spatial patterns of precipitation. In the positive phase (Figure 1d), precipitation is high over Scotland and southwestern Norway. In contrast, in the case of the reversal (Figure 1f), high amounts of precipitation were observed in the Mediterranean area and the Black Sea (see also Hurrell and van Loon, 1997). Both phases of the NAO can be described with one anomaly pattern of different sign that has to be added to the mean SLP distribution and the strength of which can be measured. Section 3, from a historical point of view and Section 6 will go into more details concerning the different North Atlantic Oscillation indices and related indices. Note that the NAOIs are usually defined as the normalised pressure difference between two stations in the vicinity of the Azores High and the Icelandic Low. The Arctic Oscillation index is defined as the first principal component time series of the SLP fields north of 20° N. There is not only a positive or negative

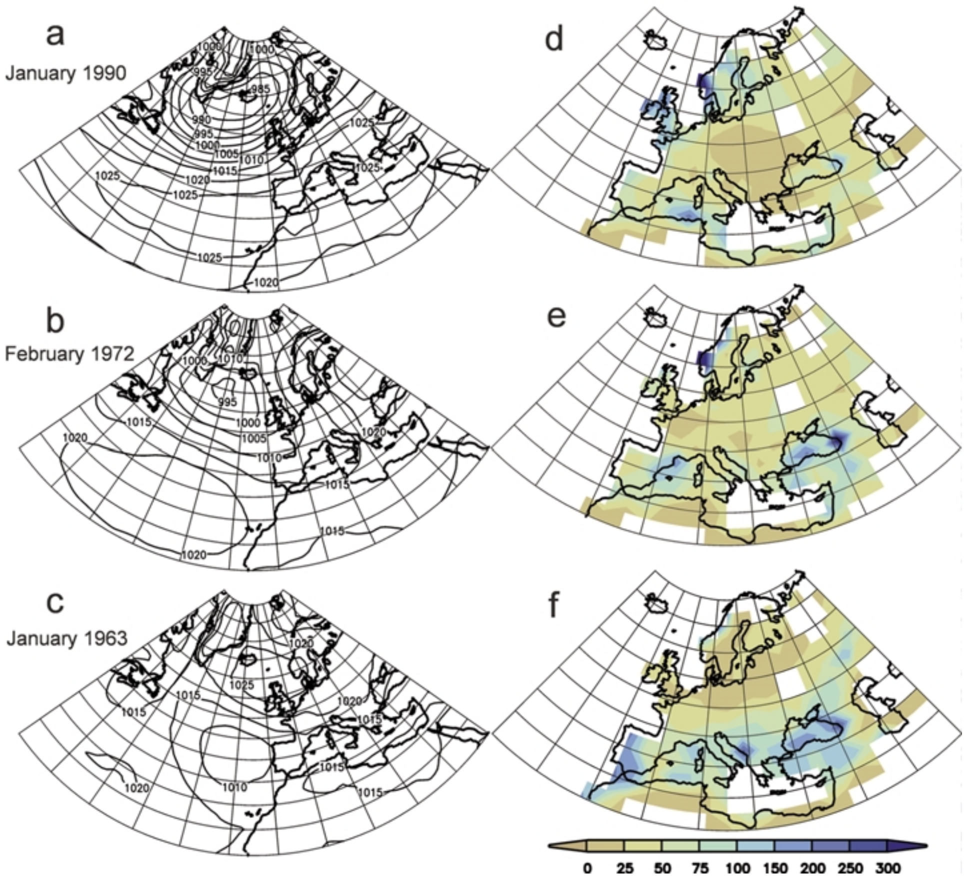

Figure 1. Monthly mean sea level pressure (a–c, hPa) and precipitation (d–f, mm) for January 1990, February 1972, and January 1963, respectively. Sea level pressure data are from NCEP/NCAR Reanalysis Data (Kalnay et al., 1996; Kistler et al., 2001), precipitation data are from Hulme (1992).

mode of the NAO, but also "everything" in between. That means, the NAO has rather a continuum of possible states than a finite set of regimes, and a bimodality in the different NAOI series is not determinable (see Section 6).

When regressing SLP time series at each grid point upon a NAOI, the resulting field of regression coefficients (often called "amplitude or regression pattern" to distinguish it from the "correlation pattern", i.e., the pattern of the correlation coefficients at each grid point) is no longer influenced by the mean SLP distribution. The dipole character of the NAO appears very clearly. This is shown in Figure 2a for a time series of winter mean SLP values (December to March) using the winter NAOI of Hurrell (1995a) from 1935 to 1999 (based on Lisbon and Stykkisholmur station pressure). The area of negative coefficients extends over the northern North Atlantic, the Arctic Basin and North Siberia and a zonally structured area of opposite sign appears over the central North Atlantic and the Mediterranean area. The

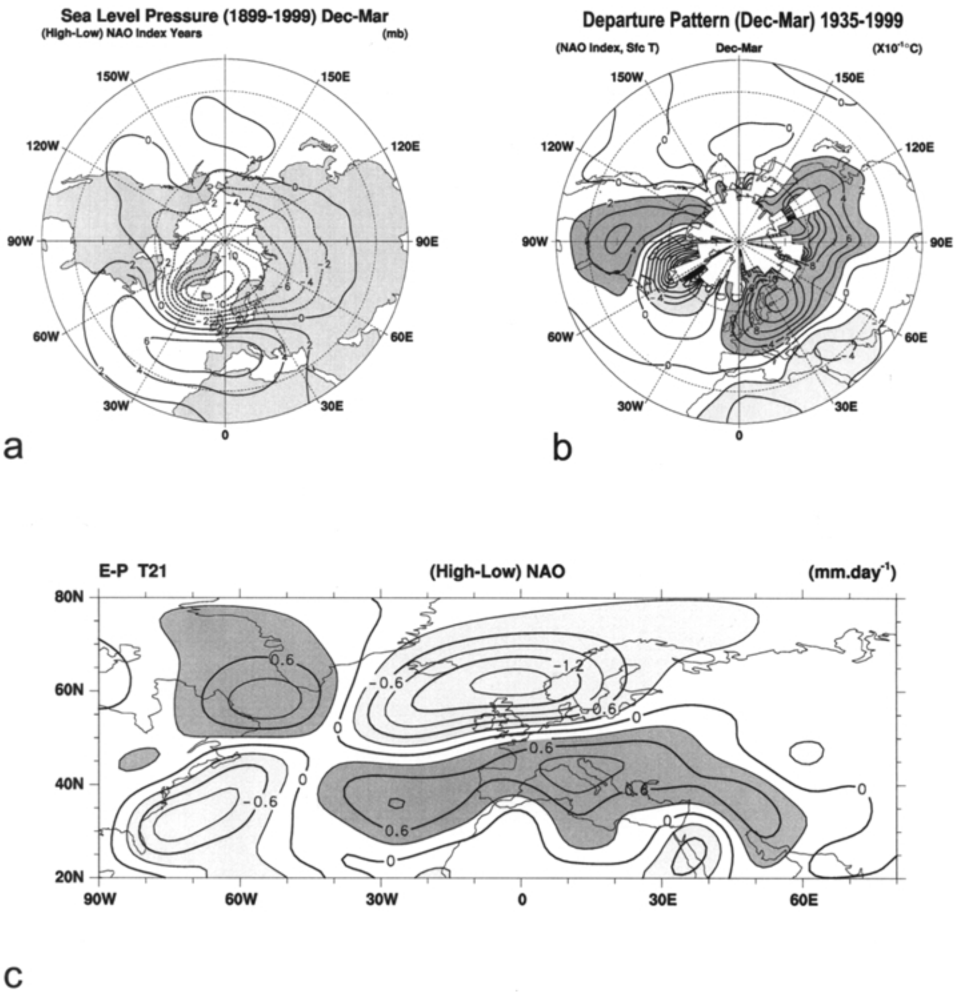

Figure 2. Spatial pattern of the NAO and its influence on temperature and precipitation, defined for winter means (December to March); (a) SPL anomalies projected on the NAOI (Hurrell, in Visbeck et al., 1998). (b) Observed surface temperature change associated with a one standard deviation of the NAOI. The regression coefficient was computed for the winters of 1935–1996. Dark (light) shading indicates positive (negative) changes (Hurrell and van Loon, 1997). (c) Precipitation anomalies associated with the NAO; E − P (Evaporation minus Precipitation) is plotted, computed as a residual of the atmospheric moisture budget using ECMWF global analyses, for high NAO index minus low index winters (after Hurrell, 1995).

highest pressure changes with changing NAOI are found over the Denmark Strait and the area east and west of the Iberian Peninsula, and not exactly over Iceland and the Azores.

Amplitude or correlation patterns for the different NAOIs and concurrent patterns look rather similar. The most important difference is that in the Arctic Oscillation amplitude pattern of SLP, the southern centre of action is more zonally elongated towards Europe and North America and the Northern Pacific centre of action is stronger, which gives a more "annular" appearance. Yet, also the AO shows a predominance of the Atlantic-European region (Deser, 2000; Ambaum et al., 2001). The subtle differences in the spatial patterns of the various indices are not shown here, the reader is referred to the figures in the papers of Thompson and Wallace (1998; 2000; 2001), Cullen et al. (2000), Deser (2000), Wallace (2000) and Ambaum et al. (2001).

By regression of the hemispheric temperature and precipitation fields on the same NAOI, we distinguish clear but different spatial patterns. The observed temperature change associated with one standard deviation change of the NAOI (Figure 2b) shows that the NAO has a dominant influence on the wintertime temperatures of the NH, especially in the area between North America and Eurasia (Hurrell and van Loon, 1997). Southwest of Iceland and east of the Gulf of Bothnia the changes are clearly higher than 1 °C per one standard deviation. Similar to the temperature changes, the precipitation anomalies related to NAOI changes (Figure 2c) show the classical seesaw between west Greenland and west Scandinavia (van Loon and Rogers, 1978; Hurrell, 1995a). This seesaw or pendulum is also well known from the winter temperatures between western Greenland and northwest Europe (van Loon and Rogers, 1978). The pendulum swings between "Greenland Above" (GA) and "Greenland Below" (GB) normal winter temperatures. During the GB-mode (an expression of the positive NAO phase), temperatures in the Greenland region and across southern Europe and the Middle East tend to be below average, but above average in the eastern United States and across northern Europe (Walker and Bliss, 1932; van Loon and Rogers, 1978; Stephenson et al., 2000). The positive NAO phase is also associated with above-normal precipitation over northern Europe (including Iceland) and Scandinavia and below-normal precipitation over southern and central Europe as well as Northwest Africa (see Figure 1). Opposite patterns of temperature and precipitation anomalies are typically found during strong negative NAO phases in the GA-mode (van Loon and Rogers, 1978). In addition, the wintertime precipitation anomalies depict the north-south gradients between Greenland and the Bermuda Islands, as well as between Scandinavia and the Mediterranean Sea.

The regional climate impact of the NAO has been subject to a vast number of detailed studies. The interested reader is referred to these studies, dealing with the correlation between the NAO and the regional or continental distribution of important variables such as temperature, precipitation, or sea-ice (e.g., van Loon and Rogers, 1978; Lamb and Peppler, 1987; Hurrell, 1995a; Hurrell, 1996; Malberg and Bökens, 1997; Rodó et al., 1997; Wanner et al., 1997; Kapala et al., 1998; Koslowski and Glaser, 1999; Osborn et al., 1999; Pozo-Vásquez et al., 2001; Slonosky and Yiou, 2001).

The NAO has distinct upper tropospheric features. Figures 3a–c show the correlations between the monthly 3-dimensional geopotential height anomaly field and a NAOI (based on station data from Ponta Delgada and Reykjavik; see Luterbacher

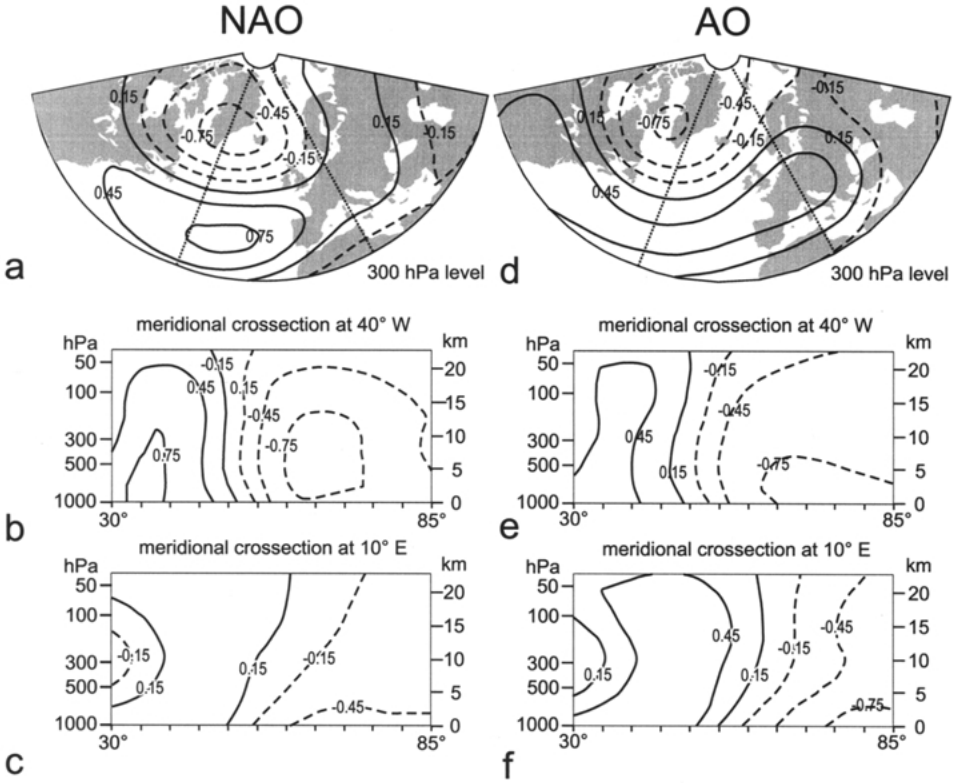

Figure 3. Correlations between monthly geopotential height anomalies and the NAOI (a–c) and the AOI (d–f), respectively, for the months November to April 1958 to 1997 (n = 240). (a, d): maps of the correlations at 300 hPa. (b, e): meridional cross-section at 40° W. (c, f): meridional cross-section at 10° E. Geopotential height data are from the NCEP/NCAR reanalysis data set (Kalnay et al., 1996; Kistler et al., 2001), AOI from Thompson and Wallace (1998). The NAOI is based on Ponta Delgada and Reykjavik station pressure (for details see Brönnimann et al., 2000).

et al., 1999) for the winter months (November to April) from 1958 to 1997. Since in this case the winter months were pooled in a sample rather than averaged, the long term mean annual cycle of geopotential height was removed prior to analysis. Correlations are displayed at the 300 hPa level (Figure 3a) and in two meridional cross-sections at 40° W (Figure 3b, representing a cut through the centres of action of the NAO) and at 10° E (Figure 3c, downwind on the European continent). The correlation pattern at 300 hPa (Figure 3a) differs from the surface patterns only in some details. The upper tropospheric structures of the NAO are displaced slightly westward. Apart from the dipole-like structure, a third feature appears at subtropical latitudes (Figures 3a and c), for which there is no corresponding surface signature, as previously noted by Appenzeller et al. (2000). The meridional cross-sections reveal a close to vertical structure over the North Atlantic but a northward displacement with height over Europe.

Figures 3d–f present the correlations of geopotential height with the Arctic Oscillation Index (AOI). In the upper troposphere and stratosphere, AO and NAO show similar structures over the North Atlantic, but geopotential height fields reveal clearly higher correlations to the AOI than to the NAOI over Europe and America (see also Thompson and Wallace, 2000). The correlations between geopotential height over the North Atlantic and the wintertime NAOI and even more so with the AOI are significant up to the 100 hPa and 50 hPa levels; thus the NAO/AO signal encompasses the whole troposphere and lower stratosphere of this region. This association is strongest in winter, which consequently, is often called the "active season". In January and February, the NAO signal was even detected in the mesopause winds at $\sim$95 km height (Jacobi and Beckmann, 1999).

The clear signal in the stratosphere merits further attention since stratosphere-troposphere coupling is relevant for many climate processes, such as solar and volcanic climate forcing, internal system dynamics such as sudden stratospheric warming, El Niño/Southern Oscillation (ENSO) and Quasi Biennial Oscillation (QBO; Perlwitz and Graf, 1995). It might also be useful for the understanding of long-term stratospheric ozone variability (Appenzeller et al., 2000; Brönnimann et al., 2000). Leading modes of stratosphere-troposphere coupled variability in the NH extracted from geopotential height fields by means of Singular Value Decomposition (SVD; Baldwin et al., 1994) or Canonical Correlation Analysis (CCA; Perlwitz and Graf, 1995) reveal a stratospheric pattern that describes strength and position of the polar vortex and a tropospheric pattern that resembles the NAO, but which exhibits a more zonally symmetric structure. As mentioned above, this pattern was later identified as "Arctic Oscillation" by Thompson and Wallace (1998; 2000; 2001) and interpreted as the surface signature of modulations in strength of the polar vortex aloft. Correlating the SLP monthly anomalies of the NH with the leading Empirical Orthogonal Function (EOF) of the 50 hPa level gives essentially the same pattern (Deser, 2000). Thus, note that the NAO and AO/NAM are coupled with the polar vortex in the stratosphere. Over the Atlantic, a pattern similar to the NAO is also found in the difference in 1000 hPa geopotential height composites between westerly and easterly QBO phases (Baldwin et al., 2001). Section 4 will go more into the details with respect to stratosphere-troposphere coupling.

To conclude this Section on the spatial aspects of the North Atlantic climate oscillations, we note that the spatial pattern of the NAO is a pronounced dipole-like pressure anomaly over the North Atlantic and has associated patterns of anomalies of temperature and precipitation over Europe, the Atlantic and even the eastern United States. The AO pattern is very similar. However, it encompasses the entire extratropical hemisphere, but with the largest signal over the Atlantic sector. The signal of both, NAO and AO is not confined to the surface but reaches up to the middle atmosphere.

3. History and Concepts

3.1. Historical review of previous NAO-related studies

In this chapter, the history of research on NAO-related topics is reviewed. It can be seen that many of the earlier views and concepts still resound in current research on NAO and AO. For other aspects of the early work, the reader is referred to the original papers and to the overviews given by Wallace (2000) and Stephenson et al. (2000).

The occurrence of periods with mild and severe winters, of course, was clearly noted by people many centuries ago. The first written evidence related to the "NAO phenomenon" dates back to at least the eighteenth century. As described in van Loon and Rogers (1978), the missionary Hans Egede Saabye made the following observations in a diary which he kept in Greenland from 1770–1778: "In Greenland, all winters are severe, yet they are not alike. The Danes have noticed that when the winter in Denmark was severe, as we perceive it, the winter in Greenland in its manner was mild, and conversely". In his book on "Historie von Grönland", published in 1765, D. Crantz wrote about the opposition of winters in the two regions. Traders and missionaries visiting Greenland were aware of this "teleconnection" between Danish and Greenland climate, although they did not have the necessary measurements for further scientific investigation.

In the nineteenth century, when more data became available, climatologists began to study the spatial characteristics of winter temperature. Dove (1839; 1841) investigated some 60 temperature series of up to 40 year length from the NH and noted that anomalies vary more pronounced between East and West than between North and South. He noted a double opposition of the monthly or seasonal temperature anomalies of northern Europe with respect to both, North America and Siberia, and found this to be in agreement with the statement by Hans Egede Saabye. Hann (1890) then illustrated the seesaw by using 42 years of monthly mean temperatures from Jakobshavn on the west coast of Greenland (69° N, 51° W) and Vienna (48° N, 16° E) in Austria. Later studies used Oslo, Norway (60° N, 11° E) instead of Vienna. One trigger for the work of meteorologists at that time was the experience of some anomalously cold winters in Europe such as the winter of 1879/80. Teisserenc de Bort (1883) started to study the positions of large pressure centres (he defined the term "centres of action") during anomalous winters. He distinguished five types of anomalous winters according to the position of the Azores High and the Russian High and to some extent also the Icelandic Low. He suggested that surface influences (such as Eurasian snow cover) were responsible for these displacements. Petterson (1890) and Meinardus (1898) investigated the influence of the Gulf Stream on weather and climate in Western Europe. They suggested that interannual fluctuations in the Gulf Stream system could be responsible for anomalous winters, and they noted that these fluctuations could affect the weather in western Iceland and Greenland in the opposite way than in Europe.

The focus of most meteorologists and climatologists at that time, however, was more concerned with describing the phenomena rather than understanding its underlying processes. Around the turn of the century, meteorologists started to think about seasonal forecasts and used a more statistical approach to study weather and climate. Hildebrandsson (1897) plotted pressure series from different sites and found a distinct inverse relation between the pressure at Iceland and the Azores. He also noted that series from the Azores and Siberia ran "parallel", whereas Alaska and Siberia showed an opposite behaviour. An impetus to this kind of "statistical" climatology was the concept of correlation, which was invented in 1877 by Francis Galton, but first published in 1888 (cf. Johnson and Katz, 1998). Walker (1909) and Exner (1913) were some of the first to apply this technique in climatology. While the focus of Walker's work was the prediction of the Indian Monsoon and the flooding of the Nile, Exner's interest was mainly the extratropical NH. He published a map of the correlation between the monthly pressure anomalies at the North Pole (approximated by the mean of three series from Greenland, northern Norway, and Northern Siberia, respectively) and some 50 sites in the NH (Figure 4a). He pointed to the annular appearance of the pattern and the strong signature in the North Atlantic and Mediterranean areas. In fact, his correlation pattern is similar to the one of the AO (Wallace, 2000). On the other hand, his work remained statistical and did not really address the physical processes behind the polar vortex.

Figure 4b shows Exner's map of correlations of pressure anomalies between Stykkisholmur and some 70 sites (Exner, 1924). This pattern resembles more the classical NAO pattern. Walker (1923) expanded his earlier statistical work to include also the North Atlantic area and addressed the "Iceland-Azores Oscillation". In his milestone paper (Walker, 1924), he described three modes which dominate world weather and introduced the terms North Atlantic Oscillation, North Pacific Oscillation (the two Northern Oscillations) and Southern Oscillation. He vaguely related the NAO to the Gulf Stream and the sea-ice dynamics in the North Atlantic, but he was sceptical about periodicities of 2 and 4.5 years of the sea-ice extent off Iceland and Iceland pressure that were discussed by other scientists at that time. Part of Walker's genius in isolating dominant modes was the way he expertly "rejected" spurious correlations using significance testing.

In the same year, Defant (1924) published a study on the monthly SLP anomaly fields over the North Atlantic. He distinguished two pairs (four types) of anomalies, where the first pair (83% of all months) corresponds to the NAO-pattern and the second pair to a strong anomaly at 55° N and a weak opposite anomaly between 10° and 30° N. By subjectively attributing to each month an anomaly type and a strength and applying a weighting procedure he was able to draw annual time series (Figure 5). He related this to the anomalies of the zonally averaged meridional pressure gradient over the Atlantic, volcanic eruptions, the zonal pressure gradient between Northern Europe and the northern North Atlantic, and the sea ice extent off Iceland. He considered these variations to be internal oscillations of the climate system of 3–5 years periodicity that were disturbed by volcanic eruptions. Defant

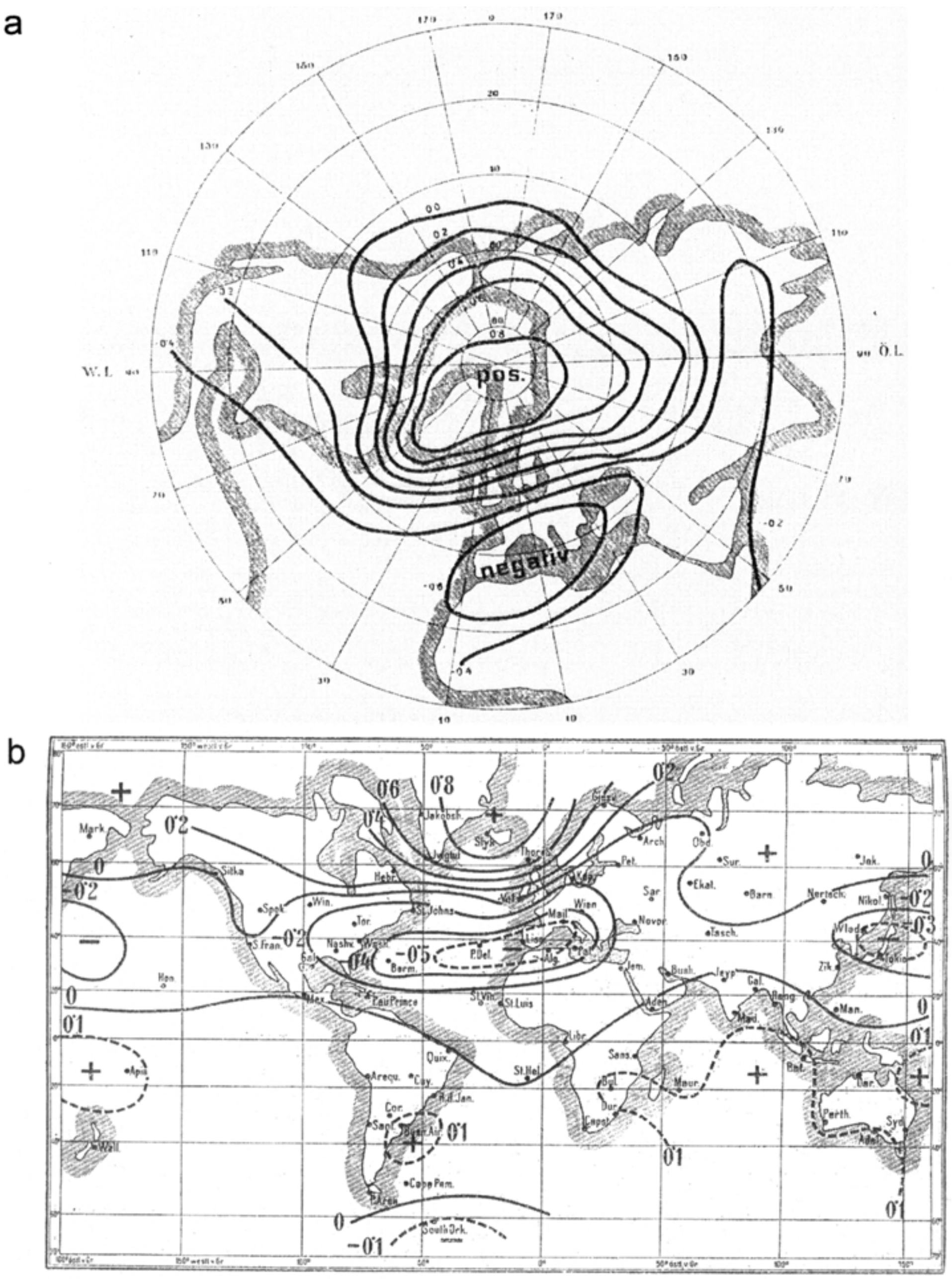

Figure 4. (a) Map of the correlation between monthly anomalies of "polar pressure" (average of three stations in northern Greenland, northern Norway, and northern Siberia, respectively) and pressure at around 50 sites of the Northern Hemisphere from 1887 to 1906 (from Exner, 1913). (b) Map of the correlation between monthly anomalies of pressure at Stykkisholmur and pressure at about 70 sites for winter months (September to March) from 1887 to 1916 (from Exner, 1924).

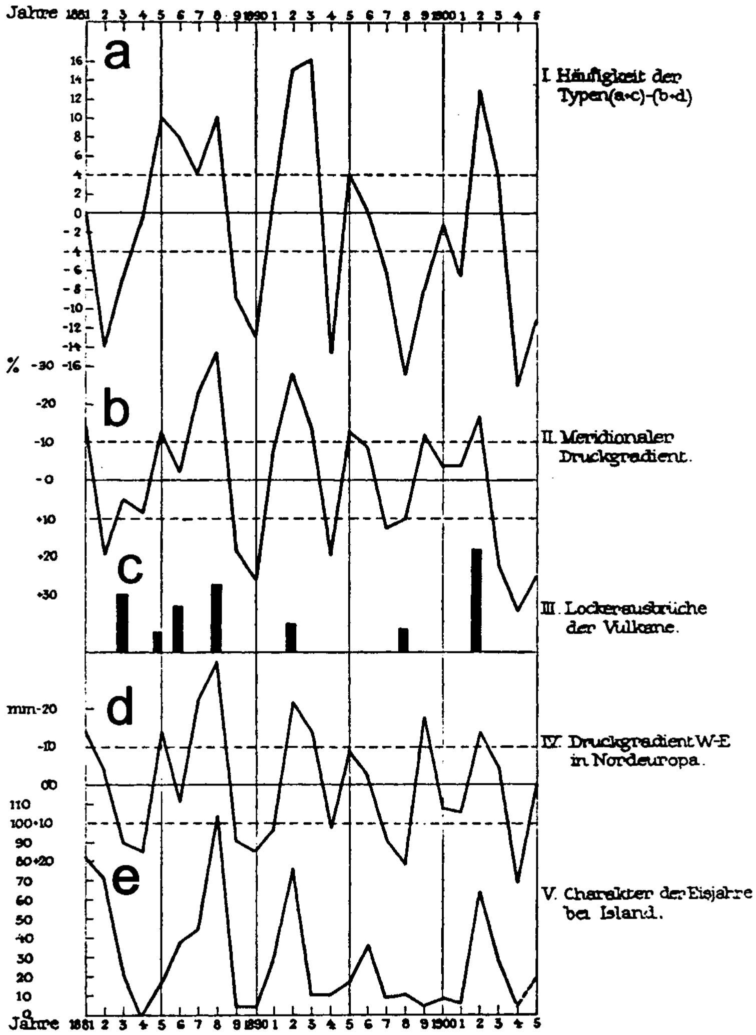

Figure 5. Time series from 1881 to 1905 of different variables from the study of Defant (1924). (a) Defant's circulation index, where negative values correspond to a positive NAO; (b) percent anomaly of the pressure gradient between 60° to 70° N and 25° to 35° N, averaged from 10° to 60° W; (c) index of volcanic eruptions; (d) pressure difference between Northern Europe (0° to 40° E, 60° to 75° N) and the North Atlantic between 60° and 70° N (in mm Hg); (e) duration of sea ice cover off Iceland (in eights of a month).

(1924) also pointed to possible relations between the North Atlantic climate and the "heat engine" of the tropical Atlantic, taking up older, speculative ideas of Shaw (1905) and Hann (1906).

Walker's concept of the NAO became very popular among contemporary meteorologists and created the need for a quantitative measure of the strength of the NAO. Walker and Bliss (1932) constructed the first NAOI in a rather complex, iterative procedure involving seven time series of temperature and SLP data from Europe and North America, by using the following formula:

$$P_{\text{Vienna}} + T_{\text{Bodö}} + T_{\text{Stornoway}} + 0.7P_{\text{Bermuda}} - P_{\text{Stykkisholmur}} - P_{\text{Ivigtut}} - 0.7T_{\text{Godthaab}} + 0.7(T_{\text{Hatteras}} + T_{\text{Washington}})/2.$$

P stands for air pressure and T for air temperature averaged over the winter period December to February. The individual series were standardised to have standard deviations of $\sqrt{20}$, and the weights were discovered iteratively. It is interesting to note that the series of Azores pressure was found to be a bad NAOI predictor and was excluded in the procedure. According to Wallace (2000), Walker and Bliss (1932) procedure can be considered to be an iterative approximation to Principal Component Analysis (PCA).

Further descriptive studies were published in the 1930s on the temperature seesaw between Northern Europe and Greenland (Angström, 1935; Loewe, 1937). At about the same time, along with a change-over from a descriptive, statistical view of climatology and meteorology to an explanatory, dynamical approach, new concepts on climate variability in the Atlantic-European region were developed. A number of theoretically motivated studies about the interaction of the zonal circulation and pressure centres were published by a group of leading meteorologists such as Rossby, Willett, Namias, Lorenz and others (Lorenz, 1967). Although the improvement of forecasts motivated the background for this research, these authors focussed more on the dynamics of the system governing equations.

Rossby et al. (1939) studied the structure and dynamics of the planetary waves in the presence of disturbances and deduced an influence of the strength of the zonal circulation on the temporal behaviour of the quasi-stationary centres of action. To address the hemispheric zonal circulation in the presence of embedded eddy disturbances, they introduced a "zonal index" defined as the zonally averaged difference between SLP at 55° N and 35° N. For Rossby, it was clear that this was a measure for the strength of the polar vortex in the free atmosphere to the north. Later, Rossby and Willet (1948) intensified their studies on the polar vortex and addressed the issue of stratosphere-troposphere coupling. Rossby's zonal index became popular for a certain time, and climatological studies of the zonal circulation were performed. Namias (1950), with a clear focus towards the improvement of forecasts, recognised the importance of latitudinal shifts in the zonal mean zonal wind. Lorenz (1950) studied the variability of the zonal mean circulation and the oscillations in the distribution of atmospheric mass. He introduced a new index, defined as the zonal mean meridional pressure gradient at 55° N. In fact, there were many different zonal indices used at that time (Kutzbach, 1970), however, none of them became popular (see Wallace, 2000, for a more detailed discussion).

Building on the impressive study by Helland-Hansen and Nansen (1920), Bjerknes (1964) reviewed ocean-atmosphere-interaction in relation to North Atlantic climate variability. He pointed to the important role of the atmosphere for the heat exchange and discussed in detail trends and anomalies in Atlantic sea surface temperatures (SST) as induced by atmospheric stresses and oceanic circulation. Bjerknes (1964) used the pressure difference between Iceland and the Azores as a simple measure of westerly flow strength. This is in fact a simple North Atlantic Oscillation index, although Bjerknes preferred the term "zonal index". A noteworthy element of this work is the dynamical analysis of past climate variability (largely based on studies done by Lamb and Johnson, 1959; 1961). Bjerknes presented ideas on the extraordinary situation in the North Atlantic from 1780 to 1820 with respect to air–sea interaction.

A much used descriptive statistical approach for isolating maximum variance patterns in the large scale circulation is the PCA of the SLP field. Although some work had been done in the 1950s (e.g., Lorenz, 1951), it was mainly Kutzbach (1970) who pioneered the use of this method for studying large scale circulation anomalies. PCA gives patterns and corresponding time series (principal components) of the patterns, which can be related to other time series. Many later studies have successfully used the same approach for SLP as well as for geopotential height fields: e.g., Trenberth and Paolino (1980), Wallace and Gutzler (1981), Barnston and Livezey (1987), Kushnir and Wallace (1989), Cayan (1992a, b), Thompson and Wallace (1998) and Volodin and Galin (1999). The first principal component pattern in winter looks relatively similar in all studies. The differences are mainly due to the selection, density and weighting of grid-points as well as due to rotated or non-rotated principal components.

The intention of these studies of the 1980s was to find the dominant modes of the low-frequency atmospheric circulation. Wallace and Gutzler (1981) pointed to the zonally symmetric, global-scale seesaw between polar and temperate latitudes in the SLP, as well as to the more regional-scale pattern resembling the so-called Pacific North American (PNA) and Western Atlantic (WA) pressure patterns at mid-tropospheric levels. By applying similar techniques to a 700 hPa level geopotential height data set, Barnston and Livezey (1987) showed that the NAO is the only low-frequency circulation pattern which is found in every month of the year.

Other studies in the 1970s re-examined the winter temperature see-saw between Greenland and Northern Europe (van Loon and Rogers, 1978; Rogers and van Loon, 1979; Meehl and van Loon, 1979). They found significant correlations between circulation and SSTs and investigated the teleconnections with the Pacific region and with the tropical climate system. These studies significantly "shaped" the current concept of the NAO as a large-scale climate mode in the North Atlantic region with important impacts in European climate. The new efforts in the field of North Atlantic climate variability in the 1970s and 1980s were partly triggered by the dominance of the positive NAO phase at that time and by the increasing requests for seasonal climate forecasts. This view is represented by Lamb and Peppler (1987) who outlined the main concept of NAO and applied it to a regional climate problem, namely to the interannual variation of Moroccan precipitation. The idea of a two point NAOI capturing the two quasi-permanent pressure centres of the North Atlantic, the Azores High and Icelandic Low, was re-introduced by

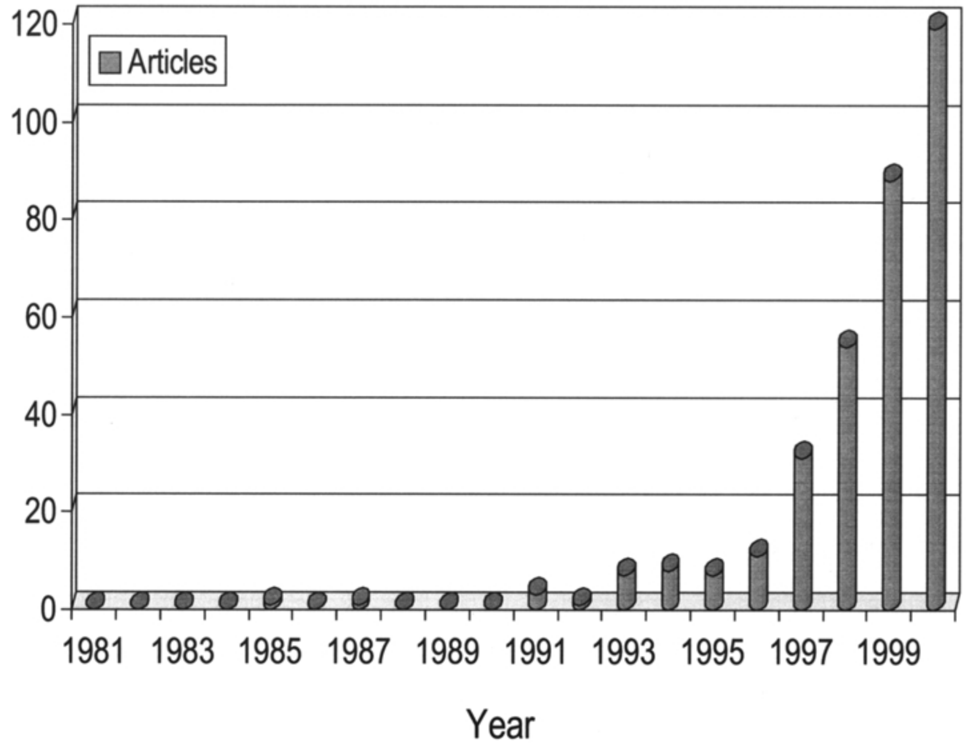

Figure 6. Bar-plot showing the increasing number of published articles containing the expression "North Atlantic Oscillation" either in the title or in the abstract during the period 1981 and 2000. Source: web of science bibliographic database.

Rogers (1984). He defined the NAOI as the difference in the standardised SLP series from Ponta Delgada, Azores minus Reykjavik, Iceland.

In the beginning of the 1990s, the NAO was studied in more detail in the light of ocean-atmosphere interactions. The NAOI was related to interdecadal variabilities of latent and sensible heat flux anomalies of the North Atlantic and the oceanic circulation (Cayan, 1992a, b; Deser and Blackmon, 1993; Kushnir, 1994). At about the same time, climate modellers began to search and study the NAO in their model simulated climates (Delworth et al., 1993). In the 1990s, the number of scientific papers on the NAO grew very rapidly. Figure 6 shows the number of all articles with the expression "North Atlantic Oscillation" in either the title or the abstracts between 1981 and 2000. The interest in NAO appeared at the beginning of 1980s (in 1984 Rogers published his NAO paper). In the following decade, the tremendous increase – a small NAO-boom – of the number of published papers indicates the growing interest on this topic, especially after the publication of Hurrell in 1995 (Hurrell 1995a). In 2000, have been published 119 papers about NAO itself, its influence and correlation with other phenomena.

Hurrell (1995a) investigated the influence of the NAO on temperature and pressure variability over the European continent on the interannual to decadal timescale. Since pressure observations at Azores go back only about one century, Hurrell defined a new NAOI as the difference between the standardised station pressure series of Lisbon minus Stykkisholmur, Iceland. This index has become the most commonly used NAOI in climate research. Jones et al. (1997) further extended an instrumental NAOI back to 1821 by using station pressure observations from Gibraltar and the Reykjavik area. For studying low frequency atmospheric variability over the Atlantic-European area, it is necessary to extend the NAOI even further back into the pre-instrumental period of the Little Ice Age. Therefore, a focus of current research is to reconstruct NAOIs using early instrumental pressure, temperature and precipitation station series and environmental proxy and documentary proxy data (White et al. 1996; Cook et al., 1998; 2001; Appenzeller et al., 1998; Luterbacher, et al., 1999; Cullen et al., 2000; Luterbacher et al., 2001a, among others). Section 6 presents our new, highly resolved monthly NAOI reconstruction back to AD 1659.

The modern debate concerning the NAO-AO/NAM gave rise to a more intense reflection also on the definition of NAO. First of all, the questions were raised whether it makes sense or not just to use a two-point index with fixed locations. On the one hand, a two-point definition is simple and easy to apply. On the other hand it is obvious that more sophisticated statistical techniques allow a better spatio-temporal description of the phenomenon. By using EOF techniques, Barnston and Livezey (1987) and Glowienka-Hense (1990) have already indicated that the two nodes of the North Atlantic dipole displace with changing seasons. Wanner et al. (1997) and Portis et al. (2001) have studied this seasonality of the NAO. They showed that both nodes migrate westward during summer. Recently, Cullen et al. (2000) and Luterbacher (2001a) tried to optimise the NAO reconstructions by using multiproxy data (see also Section 6.1). Finally, Slonosky and Yiou (2001) tried to evade the problem of the seasonal shift of the pressure centres by using two different two-point indices (Ponta Delgada – Reykjavik in summer and Gibraltar – Reykjavik in winter).

To complete the picture, it must be mentioned that Schlesinger and Ramankutty (1994) as well as Schlesinger et al. (2000) described an oscillation in the global climate system with a period of 65–70 years. Kerr (2000), by discussing the climate swing between a cold and warm phase in the North Atlantic, called this phenomenon Atlantic Multidecadal Oscillation (AMO).

What does the history of research on North Atlantic climate variability tell us? Firstly, different motivations can be distinguished, recurring from time to time: diagnostic analysis of recent climate variability, interest in the processes governing the climate system and its internal variability, the hope to be able to do seasonal forecasts and – a relatively new motivation – the analysis of low-frequency natural (as opposed to anthropogenic) climate variability using proxies and climate models. Secondly, the concepts of NAO and AO are closely linked in a historical view.

They did not develop separately from each other, nor is the NAO concept a "historical accident" (Thompson and Wallace, 1998; Deser, 2000) and the AO concept a new invention. Both concepts have roots in the historical debate. Thirdly, ever since Hildebrandson and Walker, a weak association between the North Atlantic and the Pacific SLP was admitted but the North Atlantic area was mostly considered to dominate the NH teleconnections. Fourthly, the underlying dynamical explanations for the NAO-AO oscillations still provide a major challenge to be addressed for understanding of climate variability.

3.2. North Atlantic Oscillation versus Arctic Oscillation or Northern Hemispheric Annular Mode

In their EOF analysis, Slonosky et al. (1997) pointed to the association between sea ice cover and air pressure variations. They remarked on the barotropic nature of this oscillation. Thompson and Wallace (1998; 2000; 2001) determined the leading EOF of monthly SLP fields north of 20° N weighted by area. The pattern they found has striking resemblances to the NAO in the Atlantic sector, but is zonally more symmetric with one centre of action over the Arctic region and an annular structure with opposite sign at mid-latitudes. They called this pattern "Arctic Oscillation" in order to emphasise its Arctic aspect and to show its resemblance to the dominant "Antarctic Oscillation" mode of the extratropical Southern Hemisphere (SH) (Thompson et al., 2000). More recently, it is now referred to as the Northern Hemisphere annular mode (NAM; Wallace, 2000).

What distinguishes the AO/NAM from the NAO? It is not so much the spatial pattern but the interpretation. In the NAO framework, the pressure distribution (mainly at the surface level) over the Atlantic, i.e., the Azores High and the Icelandic Low, is the main actor. The zonal mean signal evident in statistics is merely an imprint of the Atlantic. As a consequence, possible driving factors are looked for in the North Atlantic region that involve the ocean and sea-ice dynamics. In the more general AO view, the main actor is the zonal circulation and the Atlantic signal is a regional modification of the primary zonal signal. The equivalent barotropic structure and the dynamics of the higher atmosphere, namely the polar vortex appears as a more tempting forcing in this framework. According to Deser (2000), the annular appearance of the AO is caused by the Arctic centre of action, while there is no coordinated behaviour of the Atlantic and Pacific centres of action. Ambaum et al. (2001) show that the NAO reflects the correlations between the surface pressure variability at all of its centres of action whereas this is not the case for the AO. Monthly mean SLP in the Pacific and Atlantic are not significantly correlated yet both locations have large loading values in the AO pattern. The only significant correlation between centres of action in the AO pattern is between the Iceland and the Azores (Ambaum et al., 2001). In this sense, the authors conclude that the latter is not a covariance structure and state the possible interpretation that the AO is a non-local artefact of PCA.

To learn more about the differences between NAO and AO, it could be worthwhile to study the anomaly during a month where the different NAOIs and the AOI were far apart. Figure 7 shows the SLP field (raw in Figure 7a and anomalies in Figure 7b) from the 1961 to 1990 climatology of such a month (August 1976). The Hurrell-NAOI (Lisbon minus Stykkisholmur) was -2.7, a gridded NAOI (65° N/20° W–60° N/15° W minus 40° N/30° W–35° N/25° W) was -4.5, the Jones-NAOI (Gibraltar minus Reykjavik) was -3.5, and the NAOI based on Ponta Delgada and Reykjavik (Luterbacher et al., 1999) was -4.2. While there are already considerable discrepancies within the different NAOIs for this particular month, they nevertheless all agree that the meridional pressure gradient over the Atlantic was strongly reduced. In contrast, the AOI has a clearly positive value ($+0.9$). How can this be explained? If one considers the NAO as a regional phenomenon of the North Atlantic and the AO as a hemispheric zonal pattern, one would assume a major anomaly in a region other than the North Atlantic. However, as can be seen in Figure 7b, this is not the case. There is the signature of a strong polar vortex, but there is also a negative anomaly at the Azores region. The most remarkable feature is the pronounced high pressure anomaly over the North Sea, from Ireland to Norway.

The North Atlantic or the Arctic Oscillation are currently open to debate on interpretation (Deser, 2000; Wallace, 2000; Ambaum et al., 2001). On a long term, we are certain that this debate is likely to prove fruitful for our understanding of the climate system. However, in the current discussion the question has to be asked: Is the NAO obsolete? We do not believe that the NAO is obsolete. The AO is a statistical construct, resulting from an EOF analysis (Ambaum et al., 2001). If it were to be connected to one distinct mechanism, then the assumption would be that the effect of different processes on the SLP anomaly field would be strictly independent (orthogonal), which is not reasonable. The AOI is a statistically derived measure. Mainly because of its size, it explains more of the northern hemispheric temperature variability than the NAOI. The AOI captures the largest fraction of the variability of the SLP distribution and has stronger trends correlated with those in many environmental phenomena than does the NAOI. But is the AO therefore closer to the "driving mechanisms" than the NAO? Not necessarily. In contrast to Wallace (2000), we do not believe that one has to decide between the two. Rather, the AO concept has opened the eyes of many researchers not to neglect the upper troposphere and stratosphere and to consider the hemispheric-scale zonal circulation and the energetics and dynamics behind it, whereas the NAO concept has made scientists think about ocean-atmosphere interaction and search causes for low-frequency climate variability.

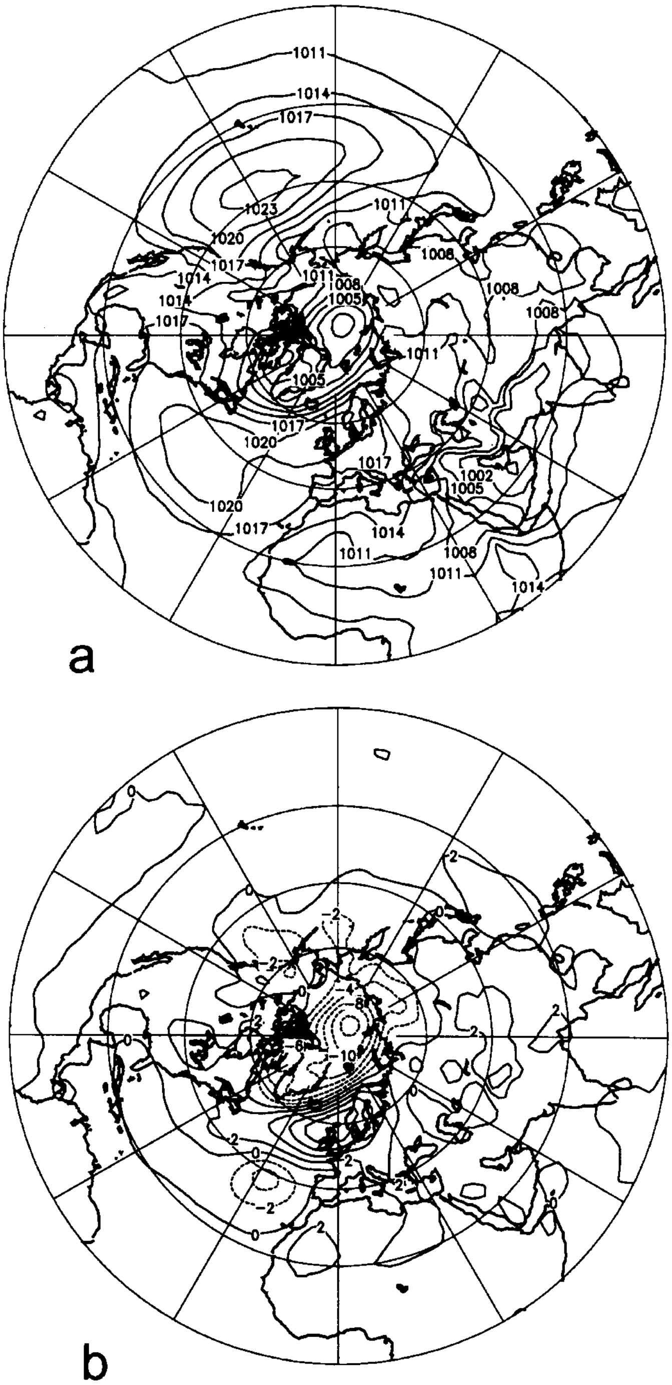

Figure 7. The extratropical northern hemispheric SLP distribution. (a) raw data, (b) anomaly with respect to 1961 to 1990, in August 1976.

4. Dynamical Aspects of the NAO

This Section will discuss different possible mechanisms for the NAO. Firstly, the question is asked whether the clue could be found simply within the atmosphere or not. Secondly, the influence of the ocean, including the important dynamical processes related to ocean-atmosphere interactions within the Atlantic region is addressed. Aspects concerning the influence of the thermohaline circulation (THC) and the SST variability in the tropical Atlantic area are also discussed in this Subsection. A third Subsection, centred on the freshwater problem in the subpolar and polar area, brings the polar sea ice, the evaporation to precipitation relationship and the run off from the huge northern landmasses to the Arctic Ocean basin into play. The pioneering paper on SST-SLP-interactions was published by Bjerknes (1964). By analysing SST and SLP anomalies, he came to the conclusion that interannual SST variability is mainly driven by changes in the heat flux from the atmosphere, but that ocean dynamics (especially heat transport) play a major role in controlling SSTs on interdecadal timescales.

In general, it is important to emphasise that many of the discussed problems are still very controversial, and open to many debates. Therefore, the crucial question about how far the NAO is an expression of a climate regime (Palmer, 1999), which itself is the result of a single climate forcing, or a certain mix or cocktail of different (natural and/or anthropogenic) forcing factors, is not tackled in this review article.

4.1. The NAO, a Purely Atmospheric Phenomenon?

The question arises whether the NAO could be a natural unforced atmospheric phenomenon or not. James and James (1989) found a long-term mode being able to create low-frequency variability, which is exclusively based on non-linear feedbacks within the atmosphere. Furthermore, Barnett (1985) and Marshall et al. (1997) showed that it is possible to reproduce NAO-like fluctuations with atmospheric General Circulation Models (GCM), which are forced with temporally non-varying SSTs.

The observational preoccupation with North Atlantic weather and climate provokes the old question as to whether the existence and the spatial arrangement of the NAO is to some extent an aggregated average of atmospheric synoptic behaviour consisting of east-northeast moving low pressure systems within the quasistationary wave in the lee of the Rocky Mountains and North America. From the viewpoint of the NAO being a regional manifestation of the AO, the planetary-wave signature embedded in the AO is induced by horizontal temperature advection as a consequence of the zonal mean flow perturbation and the strong local sources and sinks of heat, i.e., land-sea contrasts (Thompson and Wallace, 1998; 2000; 2001). In this context, it would be especially interesting to estimate the dynamical influence of Greenland, for example in two numerical experiments with and without considering the dynamical effects of the Greenland land mass

(K. Fraedrich, pers. communication). Furthermore, Hurrell (1995b) as well as Limpasuvan and Hartmann (1999) discussed the role transient and stationary eddy fluxes play for the maintenance of the different phases of the NAO-AO/NAM phenomenon. De Weaver and Nigam (2000) showed that the interactions between the zonal-mean flow anomalies and the climatological eddies make the dominant contribution to the maintenance of the NAO stationary waves.

One hypothesis that is presently discussed relates to the coupling between tropospheric and stratospheric circulation. Perlwitz and Graf (1995) as well as Kodera et al. (1996; 1999) and a large number of recent references point to the statistical connection between the strength of the stratospheric winter vortex and the tropospheric circulation over the North Atlantic. During winters with an anomalously strong stratospheric polar vortex, the NAO tends to be in its positive phase. The key actors in this mechanism are vertically propagating planetary waves originating from the troposphere and the zonal circulation at the tropopause level and in the stratosphere. At some altitude these waves break, their energy dissipates and interacts with the mean flow. Following Perlwitz and Graf (1995), the lower stratospheric zonal wind and its vertical shear, influence upward propagation of planetary waves.

It is generally assumed that the strong westerly vortex in winter provides a waveguide for efficient upward propagation of tropospheric waves and therefore, an association between the stratosphere and the troposphere ("active season"). Following Kodera and Kuroda (2000), the circulation can be forced in an AO-like way by downward propagation of zonal-mean zonal wind anomalies from the stratosphere (typically in February, March) as well as by tropospheric waves (in early winter). The slow oscillation of the zonal-mean zonal wind in the stratosphere, the Polar-night Jet Oscillation (PJO) acts as a preconditioner and feedback. Kodera et al. (1999) also point to the interesting fact that, prior to the early 1970s, the NAO and Polar Night Jet (PNJ) indices varies almost independently, while recently they exhibit a similar variability.

Stratosphere-troposphere coupling is not restricted to short time scales. Perlwitz et al. (2000) studied the dynamics of the polar vortex in a long-term integration of a climate model and found two quasi-stable climate modes that can be described as a weak vortex and a strong vortex state. Both modes are related to different stratosphere–troposphere coupling mechanisms. The climate of the second half of the twentieth century, in this framework, showed stronger similarities to the weak vortex state.

The coupling mechanisms between the stratosphere and the troposphere offer a platform for possible explanations of stratospheric controls on climate (including NAO): QBO, solar cycle (Labitzke and van Loon, 1995), explosive volcanism (Robock and Mao, 1992; Robock, 2000), ozone depletion and Greenhouse Gas (GHG) induced global warming (Graf et al., 1998). Although it seems that the atmosphere reacts in an AO-like way to some of these forcings, no final agreement is reached as to what extent the stratosphere is actively controlling the long-term behaviour of the NAO or AO (Baldwin and Dunkerton, 1999; Kuroda and Kodera, 1999; Thompson and Wallace, 2000; 2001; Thompson et al., 2000). Other scientists, however, do not believe in a stratospheric control of climate variability. Despite the more and more detailed picture we have of the statistical relations between stratospheric and tropospheric circulation, the most fundamental question remains unsolved: What is the direction of cause and effect and what is the feedback (see also Deser, 2000)?

4.2. On the NAO Dynamics within the North Atlantic Basin

According to the previous Subsection, one could argue that dynamical coupling with ocean and sea-ice would not be necessary to understand NAO dynamics. Based on the findings by Bjerknes (1964) and on progress of knowledge related to the ENSO phenomenon, one has to consider dynamical coupling, especially on higher time scales. In contrast to the ENSO, which is a rather zonally arranged tropical phenomenon, the NAO system reaches from the tropical Atlantic Ocean to the polar basin with its sea-ice system. Its spatial extent is clearly smaller, and it must also be influenced by global processes outside the Atlantic area (Hurrell, 1996). This is one important reason for the complexity of the NAO and its dynamics. We are therefore still far from having a consensus about the processes being responsible for the spatio-temporal NAO variability observed on different time scales, namely in the interdecadal range.

We first try to give, in the following Subsections, a short overview on possible processes relevant for the generation of the NAO centred system variability within the North Atlantic Ocean, including the tropical Atlantic and the northern polar basin. We are not able to discuss extensively all the relevant processes being responsible for the generation of short to long term variability of the coupled ocean–sea-ice–atmosphere system in the Atlantic area. We consciously do not split the Subsections related to different time scales or between modelling and observations. We rather try to differentiate between a few dynamical phenomena or processes, which are important for the NAO dynamics, and therefore linked to specific regions.

4.2.1. *NAO Related Atmospheric Forcing of the Ocean*

If we follow Bjerknes' (1964) concept of the time scale dependency of North Atlantic air–sea interaction, we have to consider one way interactions between atmosphere and ocean first, and then to look at complex coupled modes being relevant for NAO. On interannual time scales, observational results demonstrate that atmospheric anomalies might be able to force the variations in the ocean by anomalous heat fluxes and surface mixing (Wallace et al., 1990; Zorita et al., 1992). Hasselmann (1976) and Frankignoul and Hasselmann (1977) were able to explain this active role of the atmosphere with the framework of a stochastic climate model. In this model, weather noise, usually representing white spectra, is integrated by the much slower reacting ocean, leading to spectra that are essentially red down to frequencies where the system is stabilised by some internal feedback mechanisms. Later, Frankignoul et al. (1997) have extended the stochastic climate model to decadal time-scales. However, Stephenson et al. (2000) showed that the power spectrum of the NAO has increasingly more power at the low frequencies than expected from red noise, because of long-range dependence in the time series. Häkkinen (1999), based on a 43-year ocean model simulation (1951–93), showed that the meridional overturning cell and heat transport must be driven by the NAO related heat flux. Tsimplis and Josey (2001) found that the strengthening of the NAO from the 1960s to the 1990s explains a significant proportion of the reduction in Mediterranean Sea level over this period. The link arises from the combined effects of atmospheric pressure anomalies and changes in evaporation and precipitation.

In a recent study, Eden and Jung (2000) studied how the North Atlantic circulation responds to a forcing by the NAO on interdecadal time scales. They found that the observed and modelled developments of interdecadal SST anomalies during the periods 1915–1939 and 1960–1984 against the local damping influence from the NAO can be traced back to the lagged response (10–20 years) of the North Atlantic THC and the subpolar gyre strength to interdecadal variability of the NAO. This aspect has to be further discussed when looking at coupled modes. In conclusion, the NAO influences the Atlantic Ocean by inducing substantial changes in surface wind patterns, thereby altering the heat and freshwater exchange at the ocean surface, and influencing the thermohaline and horizontal gyres circulations (GC) (Hurrell et al., 2001).

4.2.2. *Oceanic Forcing of the North Atlantic Atmosphere and NAO*

The question how North Atlantic SSTs can influence the atmosphere and therefore also the NAO structure and dynamics is a pressing one. Frankignoul (1985) was able to demonstrate that atmospheric forced SST anomalies are damped by the turbulent surface heat fluxes, thereby influencing the atmospheric boundary layer. Several authors looked at imprints of the upper ocean structure on atmospheric circulation in winter or spring (e.g., Ratcliffe and Murray, 1970; Palmer and Sun, 1985). Together with Barnston and Livezey (1987), Peng and Mysak (1993) as well as Peng et al. (1995), Palmer and Sun (1985) have shown that strongly anomalous warm SSTs near the coast of Newfoundland are significantly related to atmospheric circulation anomalies, indicating a meridional pattern with positive SLP anomalies over the North Atlantic (that is to say in the lee of the warm pool) and negative values over north-western Europe. Czaja and Frankignoul (1999) have further demonstrated that significant anomalies of the atmospheric circulation are related to previous North Atlantic SST anomalies. For instance, a signal over the northwest Labrador Sea in late spring is associated with the dominant mode of SST variability during the preceding winter, and a NAO-like signal in early winter is connected to SST anomalies east of Newfoundland and in the eastern subtropical North Atlantic during the preceding summer.

Rodwell et al. (1999) as well as Mehta et al. (2000) used long time series of SSTs and sea-ice cover to force an ensemble of Atmosphere–Ocean General Circulation Model (AOGCM) integrations with the aim to examine the predictability of the winter NAO and AO. They found that the modelled ensemble mean NAO indices correlate much better with the observed NAO indices than a typical individual integration. Rodwell et al. (1999) argue that the SST characteristics are "communicated" to the atmosphere through evaporation, precipitation and atmospheric heating processes leading to changes in temperature, precipitation and storminess over Europe. However, Bretherton and Battisti (2000) used a simple model to explain the findings by Rodwell et al. (1999) and Mehta et al. (2000) as due to the artificial way in which such atmosphere models are forced by the ocean. They forced a linear model atmosphere/ocean interaction by high-frequency atmospheric stochastic variability and demonstrated, despite the hindcast skill, that the useful predictability, associated with midlatitude SST anomalies, is limited to one or two seasons.

By using the University of California atmospheric GCM, Robertson et al. (2000) investigated the influence of the Atlantic SST anomalies on the atmospheric circulation over the North Atlantic during winter. Among other results, they found that the interannual fluctuations in the simulated NAO were significantly correlated with SST anomalies over the tropical and subtropical South Atlantic.

4.2.3. *Dynamics of the Tropical and Subtropical Atlantic and Their Relation to NAO*

During the last few years, the relations between the dynamics in the tropical and subtropical Atlantic and the NAO attached more importance. Nobre and Shukla (1996) and Black et al. (1999) have pointed to the covarying fluctuation between the Atlantic tropical SSTs and the trade winds. Together with Enfield and Mayer (1997), they showed that stronger than normal southeasterly trade winds in the SH are associated with negative SST anomalies and vice versa. Chang et al. (1997) demonstrated that the surface heat flux plays an active role in the tropical Atlantic. Nevertheless, it is questionable whether the correlation in the decadal dipole – with a seesaw between negative SST anomalies in the SH and positive SST anomalies north of the Intertropical Convergence Zone (ITCZ) – is highly significant or not (Houghton and Tourre, 1992). Several studies point to the possibility that the climate variability and the position of the ITCZ in the tropical Atlantic Ocean is sensitive to the cross-equatorial SST differences (Hastenrath and Heller, 1977; Folland et al., 1991; Servain, 1991; Chang et al., 1997). The dipole might also influence the rainfall over parts of South America (Moura and Shukla 1981) and the Sahel region (Lamb and Peppler, 1991). The variability of the NAO seems to be coupled to the northeasterly trades and the SSTs on the NH (see Figure 8). Recently, George and Saunders (2001) demonstrated that the dominant mode of wind speed variability in the wintertime tropical is represented by the NAO. They also showed that that the winter NAOI determines monthly precipitation levels

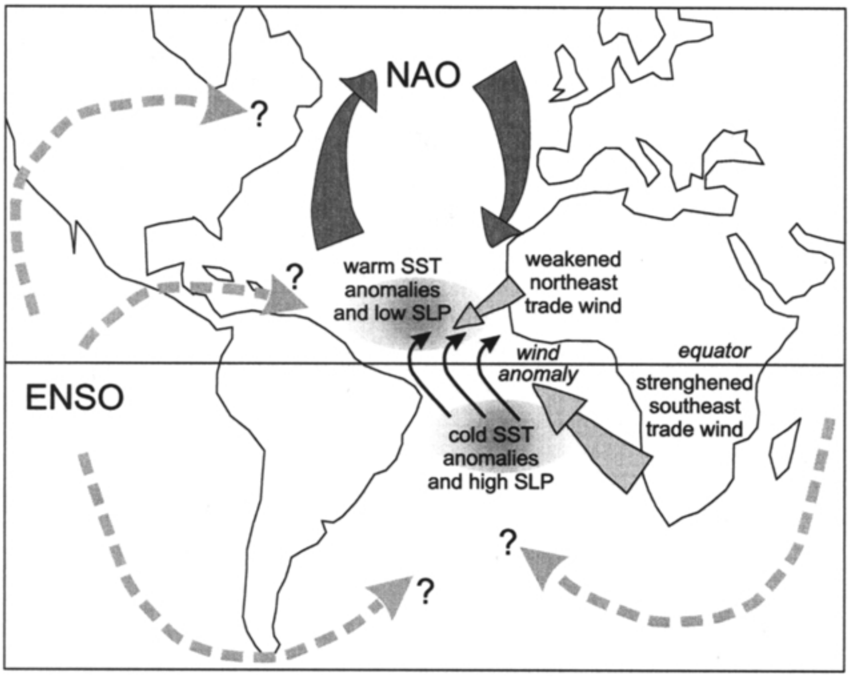

Figure 8. Influence of ENSO and important mechanisms of Tropical Atlantic Variability (after Chang, adapted from Visbeck et al., 1998).

across the northern Caribbean throughout the following year. In future, more studies need to deal with the direct coupling between the NAO and the tropical Atlantic SST-SLP-wind-precipitation system (Venegas et al., 1997; Tourre et al., 1999).

In a series of modelling experiments, Sutton et al. (2001) showed that the tripole pattern of the North Atlantic SSTs induces a significant response of the atmosphere over the tropical Atlantic as well as over the mid-to-high latitudes. The low latitude response is forced by the low latitude SST anomalies, but the high latitude response is influenced by the extratropical SST anomalies as well as those in the tropics.

4.2.4. *NAO, Thermohaline Circulation, Gyre circulation and Coupled Modes*

Bjerknes (1964) emphasised that ocean dynamics are important in controlling SSTs on interdecadal time scales. It is therefore also important to look at these dynamics as well as at coupled modes if one wants to understand longer term NAO variability.

Eden and Jung (2000) showed that the relationship between the local forcing by the NAO and observed SST anomalies on these time scales points towards the importance of these ocean dynamics for the generation of the studied SSTs. Especially for the interpretation of the decadal to multidecadal climate variability the dynamics of both, THC and GC have to be understood and considered in ocean-atmosphere feedback studies. Related to the dynamics of the interdecedal variability, Latif (1998) states that some coupled ocean–atmosphere models exhibit relative featureless red spectra in some oceanic key quantities, but other models simulate interdecadal oscillations, with statistically significant spectral peaks superimposed on the red background. Here we can only refer to a few selected publications being important for the understanding of the NAO phenomenon on the interannual to multidecadal time scale.

Cayan (1992b) concluded that the dominant atmospheric circulation modes produce basin-scale anomalies of latent and sensible heat fluxes over the Northern Oceans. Seasonally, in the extra-tropic regions, sensible and latent heat transfer from the ocean to the atmosphere are most important in autumn and winter when the near-surface vertical gradients of humidity and temperature are largest and wind speed is highest (see also Wilby et al., 1997). The dominant mode in winter is a warm pool between 40° N and 60° N and a cold pool between 20° N and 40° N, or vice versa. Further, the SSTs are negatively correlated with the overlying westerly winds, and above all, the warm pools in the northern Atlantic ocean lead to a well-marked weakening or blocking of the westerly zonal flow (Deser and Blackmon 1993; Kushnir 1994; Battisti et al., 1995). Recently, Sutton et al. (2001) investigated the feedback from the atmosphere onto the Atlantic SST tripole which is the first EOF of wintertime SST variability (Cayan, 1992a, b). They found that the expected negative feedback is significantly modified at low latitudes by the dynamical response of the atmosphere.

Deser and Blackmon (1993) demonstrated that the wintertime SSTs, the sea-ice and the atmospheric fluctuations over the subpolar gyre, change synchronously on a decadal timescale (10–15 years), and Kushnir (1994) showed that decadal changes in SST and SLP over the subpolar and subtropical Atlantic gyres mark coupled ocean- atmosphere interactions. Hansen and Bezdek (1996) as well as Molinari et al. (1997) used the sub-surface temperature variability, which is less contaminated by noise than the more stochastically forced SSTs, to study the relations between the decadal-scale ocean GC and the NAO. They generally confirmed Deser and Blackmons (1993) findings.

Latif and Barnett (1996) and Grötzner et al. (1998) showed that the sub-surface heat content anomalies in the central North Atlantic are the result of an enhanced subtropical ocean GC, which transports warmer tropical water poleward. The atmosphere responds in an anticyclonic atmospheric circulation and a weakening of the storm track activity. Another prominent explanation of the longer-term oscillatory behaviour of the NAO bases itself on the dynamics of the THC. If the THC is anomalously strong, the SSTs show positive anomalies, the NAO is strengthened, the freshwater fluxes off Newfoundland and the Greenland Sea are stronger and the negative sea surface salinity – transported by the subpolar gyre – is possibly reaching the convectively active region south of Greenland and can therefore initiate a damping of the THC. This weakening reduces the poleward heat transport and the formation of negative SST anomalies and completes the whole cycle with a duration of about 35 years (Timmermann et al., 1998). Due to the enhanced fresh water cycle with more precipitation in the North Atlantic basin, most of the recent coupled ocean-atmosphere models actually diagnose a reduction of the THC, but the magnitude varies considerably (Manabe and Stouffer, 1993; Stocker and Schmittner, 1997; Rahmstorf, 1999). In a recent publication, Delworth and Dixon (2000) described the results of experiments with a coupled ocean-atmosphere model. They showed that the above mentioned weakening of the THC could be delayed by several decades in response to a sustained upward trend in the AO/NAO during winter (as was observed during the last 30 years). The NAO+ related strong winds extract more heat from the ocean, and the cooler and denser upper ocean opposes the previously described weakening of the THC.

Other coupled model results (Delworth et al., 1993; Delworth et al., 1997) showed that the THC has an irregular oscillation with a time scale of approximately 40–60 years which involve large-scale interactions between Arctic fresh water and ice export. The irregular oscillation appears to be driven by density anomalies in the sinking region of the THC. The connected spatial patterns of SST anomalies show resemblance to a pattern of observed interdecadal variability in the North Atlantic (Kushnir, 1994; see above). These SST anomalies induce surface air-temperature anomalies over the North Atlantic, the Arctic and northwestern Europe (Delworth et al., 1993). However, the mechanisms causing a feedback of THC variations to the atmosphere are not well understood. In particular it is not known which pathways for propagation within the ocean are most relevant, and which factors determine the time scale of coupled oscillations.

An important, but difficult question is how THC, GC and NAO influence the Gulf Stream/North Atlantic Current (NAC) system (and vice versa). Over the last three decades (1966–1996), the annual mean latitudinal position of the Gulf Stream has been clearly correlated with the NAO dynamics. Positive NAO indices favour a more northern path of the Gulf Stream, with a time lag of about 2 years (Taylor and Stephens, 1998). This time lag might be associated with the adjustment time of this branch of the ocean circulation. More than half of the interannual variability in the position of the flow system can be explained. The unexplained part can be accounted for by the ENSO in the Pacific, with the Gulf Stream being displaced northwards following ENSO events. This provides a link between events in the equatorial Pacific and the circulation and weather conditions of the North Atlantic (Taylor et al., 1998; see also Chapter 4.3). Joyce et al. (2000) hypothesised that the SST signal produced by the latitudinal shifts of the Gulf Stream might influence the position of the extratropical winter storm track and thus the NAO. Recently, Frankignoul et al. (2000) investigated the lead and lag relation between the Gulf Stream position (derived from the Topex/Poseidon altimeter) and the changes in SLP, surface wind stress and SSTs for the period 1992–1998. This high NAOI period is one of unprecedented northward excursion of the Gulf Stream in the 45 years record, with the Gulf Stream 50–100 km north of its climatological mean position (Frankignoul et al., 2000). They found a dominant signal of northward (southward) displacement of the Gulf Stream axis 11 to 18 months after the NAO has reached a more positive (negative) phase. A SST warming (cooling) peaking north of the Gulf Stream is also seen to precede the latitudinal shifts (Frankignoul et al., 2000). There is evidence, that the North Atlantic SST anomalies have an influence on the NAO, but the Gulf Stream shifts seem to have no direct impact on the large-scale atmospheric circulation (Frankignoul et al., 2000).

Most recently, four other studies (Christoph et al., 2000; Häkkinen et al. 2000; Seager et al., 2000; Marshall et al., 2001) dealt with the understanding of low frequent, above all decadal scale, North Atlantic climate variability by studying coupled ocean–atmosphere phenomena. Their results again reflect the complexity of the underlying dynamical processes. Two papers (Häkkinen, 2000; Marshall et al., 2001) report about the existence of coupled modes including the influence of GC and THC. The two other (Christoph et al., 2000; Seager et al., 2000) could not find a clear indication showing that ocean transports could significantly lead or lag typical SSTs or surface fluxes. Finally, Hastenrath and Greischar (2001) analysed reanalysis data between 1958 and 1997. They showed, for the NAO-phase, that the weakened Azores High entails slower trade winds that through reduced evaporation and wind stirring are conducive to warmer sea surface and slower midlatitude westerlies that through reduced Ekman transport lead to colder waters on the poleward side of the anticyclonic axis. Similar, the weakened cyclonic circulation around the Icelandic Low makes for a warmer sea surface. These SST anomalies are then imparted to the overlying atmosphere.

4.2.5. *Influence of Arctic Ocean Circulation and Sea Ice Dynamics on NAO*

It is widely known that the dynamics within the polar basin also influence those of the North Atlantic ocean and the overlying atmosphere. The interest for these phenomena grew considerably after the period 1968–1982 when a widespread freshening of the subpolar gyre waters, the so-called Great Salinity Anomaly (GSA), was observed and analysed (Dickson, 1988; Mysak et al., 1990). Dickson et al. (1988), Aagard and Carmack (1989) as well as Walsh and Chapman (1990) pointed to the significant influence of the freshwater flux from the continents into the polar basin which can be a critical factor in determining the rate of deep convection in the high-latitude North Atlantic and influence the thermohaline circulation (Broecker et al., 1985; Aagard and Carmack, 1989; Stocker and Mysak, 1989). These freshwater fluxes are also connected with the evaporation minus precipitation amounts over northern North America and Eurasia, and therefore, with the runoff from the big rivers, such as Ob, Jenisej, Lena and McKenzie flowing into the polar basin (Peng and Mysak, 1993).

During the GSA, a large fresh water pool was observed in the northern North Atlantic Ocean (Dickson et al., 1988). Walsh and Chapman (1990) report that the THC was therefore clearly reduced. Contrarily, it was possibly intensified during the 1920s and the earlier 1960s (Deser and Blackmon, 1993). Later on, in the early 1980s, the SSTs in the North Atlantic experienced large fluctuations again, the so-called “smaller GSA” (Reverdin et al., 1997).

Several authors started to look at possible coupled modes with important feedback mechanisms. Mysak et al. (1990) described a negative feedback loop for the subpolar North Atlantic, linking Arctic cyclogenesis, precipitation, runoff, salinity, sea-ice extent, ocean stability, convective overturning, poleward oceanic heat transport and heat flux into the atmosphere in the Arctic: Periods with positive SST anomalies in the Labrador Sea are connected with higher than normal air pressure over Greenland and therefore a weaker climatological Icelandic Low. Higher air pressure over Greenland leads to a greater than normal flow of ice and fresh water through Fram Strait into the Greenland Sea (Aagard and Carmack, 1989; Power and Mysak, 1992) and therefore to a negative salinity anomaly which is advected around the subpolar gyre into the Labrador Sea. Due to the weakening of the Icelandic Low the storm tracks are diverted south of their normal position, and the NAO index is decreasing. It is particularly interesting that this link to the NAO dynamics corresponds to a decadal timescale. Wohlleben and Weaver (1995) presented a similar feedback loop, starting with deep convection in the Labrador Sea and not relying upon Greenland–Iceland–Norwegian (GIN) Seas cyclogenesis and Mackenzie River runoff to explain ice export through Fram Strait. Based on Boolean delay model calculations they proposed a time period of 21 years for one cycle of their feedback loop. Based on 40 years of sea-ice concentration (SIC) and SLP data, Mysak and Venegas (1998) presented a new feedback loop for atmosphere-ice-ocean interaction in the Arctic. It starts with a SIC anomaly in the Greenland–Barents Sea and follows – besides its strong relation to the NAO dynamics – a timescale of about one decade. In addition, Koslowski and Loewe (1994) and Koslowski and Glaser (1999) have shown that the time series of the accumulated areal ice volume in the Western Baltic Sea are negatively correlated with a temporally corresponding NAO winter index.

A series of modern studies tried to look in more detail at these mechanisms. Slonosky et al. (1997) used EOF analysis to investigate the relationship between sea-ice cover (SIC), SLP and temperature variations. The first winter SIC EOF is associated with the GSA between 1968 and 1982. The correlation between this ice EOF 1 and the atmospheric anomaly fields are highest if this EOF 1 leads the atmospheric anomaly by one year. Yi et al. (1999) found out that it is the NAO rather than the AO that is most strongly coupled to the sea-ice variability. Deser et al. (2000) studied the trends between 40 year series of SIC and reanalysis data for winter and summer. They showed that the large-scale changes of surface air temperature and SLP, which closely resemble to NAO are associated with the dominant patterns of winter sea-ice variability.

Recently, Dickson et al. (2000) studied the complex response of the Arctic to annual and longer-period changes in the NAO during winter. They concentrated on the postwar period that includes the most comprehensive instrumental record and contains the largest directly recorded low-frequency change in NAO activity. This shift was accompanied by an intensifying storm track through the Nordic Seas, a radical increase in the atmospheric moisture flux convergence and winter precipitation in the studied sector, an increase in the amount and temperature of the Atlantic water inflow to the Arctic ocean, a decrease in the late-winter extent of sea-ice throughout the European subarctic, and – at least temporarily – an increase in the annual volume flux of ice from the Fram Strait. Vinje (2001) also showed that the sea-ice extent in the Nordic Seas measured in April decreased by about 33% over the past 135 years. Nearly half of this reduction was observed in the period ~1860–1900. The correlation between the NAO scaled winter circulation and the subsequent April ice extent in the Nordic Seas was strongly negative, whereas it became positive for the Newfoundland-Labrador Sea. As mentioned earlier, climate in winter – especially in the subpolar and polar area – can not only be diagnosed based on the processes related to the leading EOF of SLP (AO/NAM or NAO). Skeje (2000) found that the second EOF north of 30° N, with its most prominent center of action over the Barents region (called "Barents Oscillation"), has a high temporal correlation with the sensible heat loss of the Nordic Seas and correlates well to Eurasian surface air temperature anomalies.

It can finally be asked, whether the spectral peak of the NAO which is observed at about 6–10 years (Hurrell and van Loon, 1997; see also Figures 10 and 13), is at least partly related to such a process. Venegas and Mysak used a frequency-domain singular value decomposition approach (Mann and Park, 1999) to analyse nearly 100 years of sea-ice and SLP data. They found four dominant signals with periods of about 6–7, 9–10, 16–20 and 30–50 years. Their study nicely revealed the complexity of the involved processes, which are strongly related to NAO and AO dynamics (especially on the 9–10 year timescale).

4.2.6. *Influence from Outside the Atlantic Basin and Possible Correlations with Other Phenomena*

The influence of anomalous tropical Atlantic SSTs on the tropical and extratropical atmosphere was shortly discussed in an earlier Subsection. It is evident that NAO dynamics can not be fully understood if the global scale energetics, especially the heat fluxes in the Pacific area, are not taken into account (Bjerknes 1966). A related interesting question concerns the correlations between the NAO (or the AO/NAM) and other teleconnections. In addition, the question is posed as to how the Pacific system is connected to the processes in the tropical Atlantic (e.g., through the influence of ENSO; Hastenrath et al., 1987).

Based on high resolution simulations with the coupled atmosphere – ocean European Centre/Hamburg Model (ECHAM/OPYC), Latif et al. (2000) and Timmermann et al. (1999) state that a warmer climate with a transition towards permanent El Niño conditions, increases the freshwater export from the Atlantic to other ocean basins. Schmittner et al. (2000) support this hypothesis by using two reanalysis data sets (Appenzeller et al., 2000). Figure 8 indicates possible links between ENSO and NAO that can be transmitted via the ocean as well as via the atmosphere. In wintertime, the PNA teleconnection pattern which is clearly associated with tropical Pacific SST anomalies in the sense that the PNA index has the tendency to be positive during warm events (Wallace and Gutzler, 1981; Kushnir and Lau, 1992; Leathers and Palecki, 1992). Figure 8 also points to the already mentioned possible coupling between the NAO and the tropical Atlantic Ocean.

Rogers (1984) investigated the association between the NAO and the ENSO in the NH. He found that both teleconnections are associated with significant SLP differences over much of the NH except for Siberia and western North America. In the 80 winters with data, simultaneous appearance of the two modes only seems to occur by chance. Fraedrich (1994) pointed to the important fact that the "noise level" may almost completely hide atmospheric teleconnections. Therefore, conditional statistics must be used to detect possible physical mechanisms. He showed that, for winter, a regional response becomes apparent in form of cyclonic Grosswetter (large-scale flow) and a more northern route of the storm track for warm ENSO events at the peak of the episodes, and anticyclonic Grosswetter with a more zonal orientation for cold events.

There is still a strong interest in the signature of the NAO at the continental scale on the pressure field, and for its statistical relation to the storm tracks. Serreze et al. (1997) presented a climatology of the Icelandic Low. They showed that the cyclonic activity during the NAO+ phase of the recent years has only increased within the region north of 60° N. Rogers (1997) investigated the association of the North Atlantic storm track variability with the NAO and the general climate variability in Northern Europe. He showed that the link of the storm tracks with the low-frequency SLP anomalies in the extreme northeastern Atlantic is clearly stronger than that with the NAO. This again shows that other modes such as the EU (Barnston and Livezey, 1987; Luterbacher et al., 1999) are also relevant for determining storm behaviour.

By using multivariate linear regression, Hurrell (1996) showed that nearly all of the wintertime cooling in the northwest Atlantic and the warming across Europe and downstream over Eurasia since the mid-1970s results from changes in the NAO, with the NAO accounting for 31% of the hemispheric interannual temperature variance in winter over 60 winters (1935–1994). By using longer NAO indices Osborn et al. (1999) show that the 31% explained variance reduces to zero over 1895–1920 and is about 25% over the period 1851–1894. Recently, Hoerling et al. (2001) presented evidence that North Atlantic climate change since 1959 is linked to a progressive warming of tropical SSTs, especially over the Indian and Pacific oceans. Therefore, the whole tropics, not just the Atlantic sector, must be considered if the North Atlantic climate variability will be understood.

In conclusion, the NAO can be considered as a phenomenon that is strongly determined by natural atmospheric processes. Anyhow, many studies dealing with atmosphere–ocean–sea-ice interaction demonstrate that one way interactions or coupled modes exist, especially on the decadal scale. Their influence might be rather weak if a certain anomaly (e.g., a SST pattern), is not extremely strong and persistent. In addition, many processes are still unknown. Even the prediction potential of NAO variability seems to be much smaller than in the case of, e.g., ENSO (Hurrell et al., 2001), there is some hope that predictability could arise from the influence of slow changes in the ocean or from external factors such as rising levels of greenhouse gases.

5. Spatial Structure of the NAO

Figure 9 attempts to visualise the spatial structure of the two (positive and negative) phases or states of the NAO with its major components of the ocean–atmosphere–sea-ice system. It was constructed based on an extended review of the available literature and on discussions with different researchers. Note that the Figure is schematic in many of the represented structures. In addition, it suggests that the two phases or states of the NAO mode are static, quasi-stable, and show an unrealistic bimodal frequency distribution. This is a strong simplification and partly wrong because the NAO is an intermittent phenomenon, has no preferred frequencies except perhaps at low frequency (Stephenson et al., 2001), and due to its intermittent character, does not reside in only two quasi-stable states (see also Figures 10 and 11).

As mentioned above, the two NAO phases or states (Figures 9a and b) are related to the often used expressions "high and low index circulation type", even though this term is normally used for higher frequency atmospheric phenomena along the mid-latitude westerlies (Namias, 1950). The following components can be distinguished:

- Two pools with anomalous ocean temperatures are situated east/southeast of Greenland and west of North Africa (both with negative anomalies in the NAO+ phase, and positive anomalies in the NAO-phase). Two other pools with contrasting temperature anomalies exist over the North American Basin and around the English Channel (Cayan, 1992a,b; Kushnir, 1994; Becker and Pauly, 1996).
- Beside the interactions with the atmosphere the four pools interact with (or are the result of) the circulation in the ocean, namely the subtropical and the subpolar gyre, the Gulf Stream, the Mid-Atlantic and West Norwegian Current (McCartney and Talley, 1984; Spall, 1996a,b; McCartney, 1997; Kerr, 1997; Sutton and Allen, 1997).
- The subtropical and mid-latitude ocean circulation system is also connected to the polar basin – namely through the Labrador Current and the Transpolar Drift Stream (Walsh and Chapman, 1990) – and the formation and distribution of sea-ice. During the NAO+ phase, the Labrador Sea is covered by sea-ice (sustaining a strong deep convection; Marshall et al., 1993), the Labrador Current extends far south and southward water and sea-ice transport along the Transpolar Drift Stream east of Greenland is decreasing. During the NAO-phase, the Labrador Sea is open (with a decreased deep convection) and the

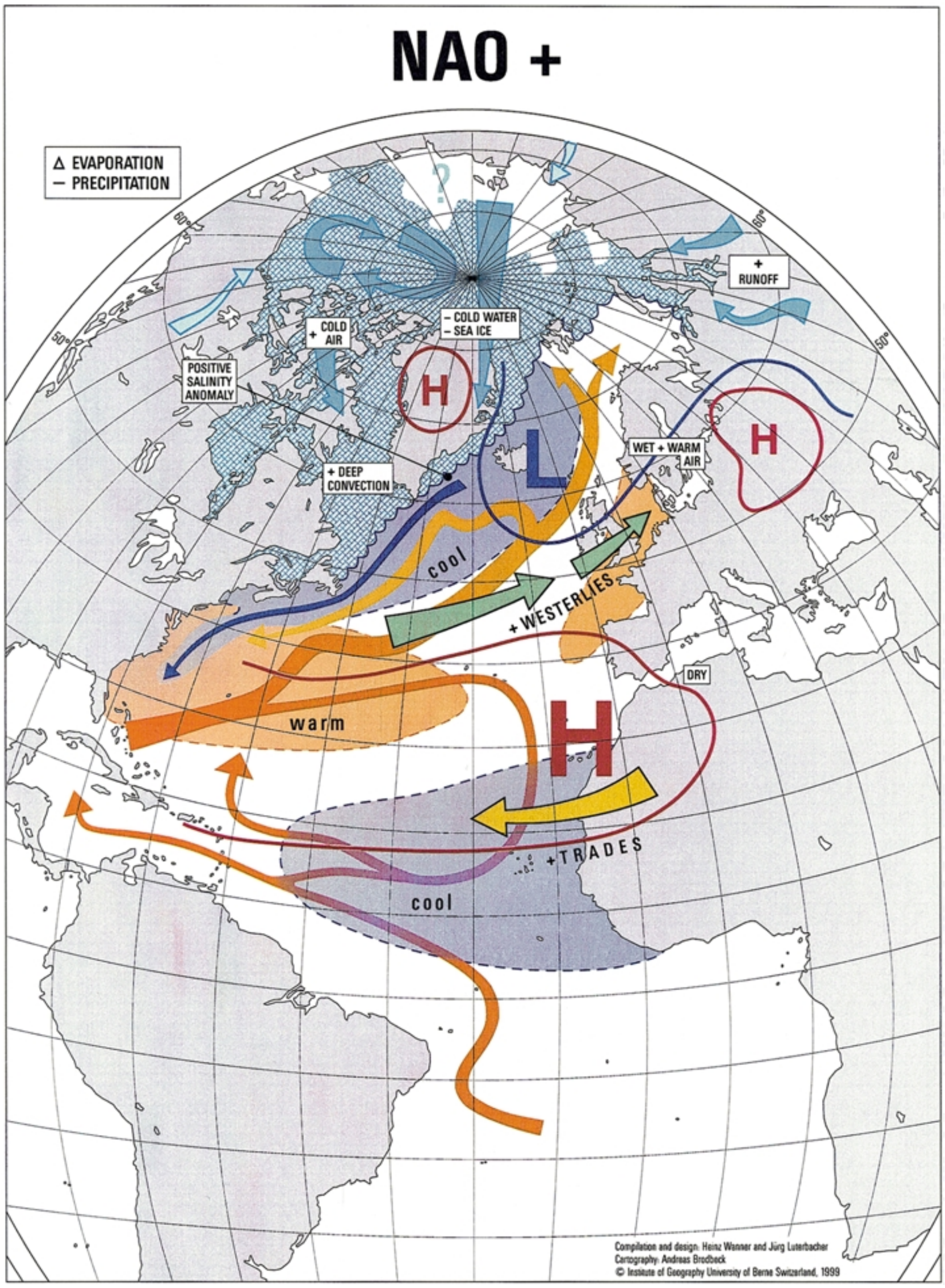

Figure 9a. Graphical representation of the two modes or states of the NAO, based on discussions with different researchers and a review of different papers. Surfaces mark SSTs and sea-ice extension, arrows show the flow systems in ocean, atmosphere and rivers, blue and red lines indicate near surface sea level pressures and white rectangles describe characteristic climate conditions or important processes. (a) Positive mode, (b) Negative mode.

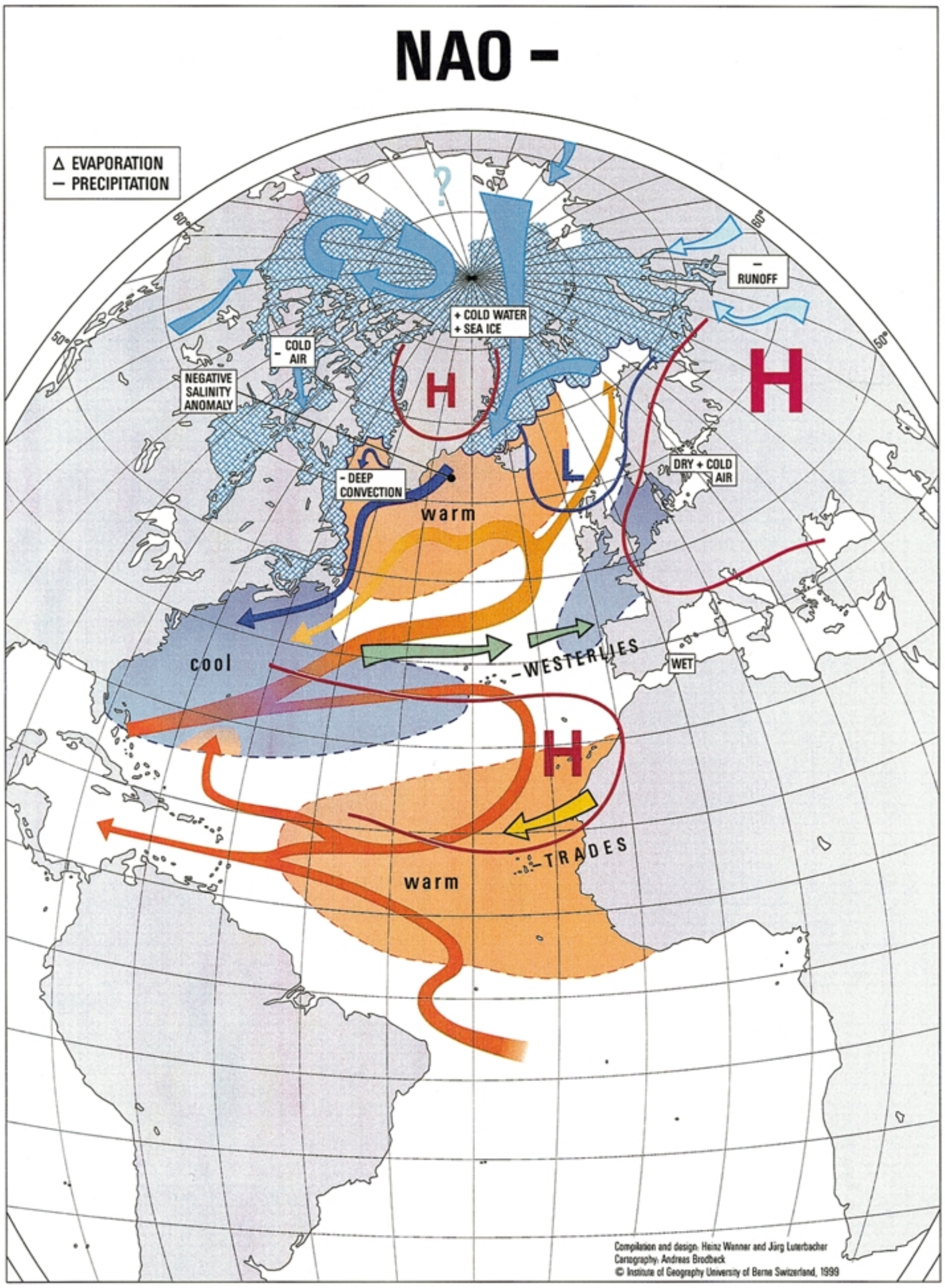
NAO -
Δ EVAPORATION
– PRECIPITATION
– RUNOFF
+ COLD WATER
+ SEA ICE
– COLD AIR
NEGATIVE SALINITY ANOMALY
H
H
L
– DEEP CONVECTION
DRY + COLD AIR
warm
cool
– WESTERLIES
WET
H
– TRADES
warm
Compilation and design: Heinz Wanner and Jürg Luterbacher
Cartography: Andreas Brodbeck
© Institute of Geography University of Berne Switzerland, 1999

Figure 9b.

sea-ice and water transport through the Fram Strait becomes stronger with an increasing NAO (Wohlleben and Weaver, 1995; Kwok and Rothrock, 1999).

– The sea-air exchange and the ocean surface currents at the border between the mid-latitudes and the polar basin are influenced by the air pressure configuration over Greenland. In the NAO+ phase, the temperature gradient between Greenland and the cool ocean surface to the south is minor. Therefore, the flow divergence over Greenland is much smaller than in the negative mode and the air pressure tends to be low which again influences the Labrador Current and the Transpolar Drift Stream.
– Energy and mass fluxes to and from the sea and ice surfaces influence the air pressure and the wind systems from the lower to the higher atmospheric levels. In the NAO+ phase, the Icelandic Low and the Azores High are well developed, with anomalous low and high pressure cores, respectively. This pressure configuration leads to strong pressure gradients and strengthens the westerlies and the trade winds. The axis of the westerlies orients southwest–northeast and shifts to the north. In the negative phase, the opposite effects are observed, with the axis of the westerlies clearly zonal and lying further south.
– The described configuration of the pressure systems influences the storm tracks (Lau, 1988; Rogers, 1997; Serreze et al., 1997) and, therefore, the transport of heat and moisture to the European continent (Hurrell, 1995a; Halpert and Bell, 1997). In the NAO+ phase with a more northern jet axis, wet and warm winter climate is observed in Scandinavia, while cool and dry conditions predominate in southern Europe and northern North Africa. Almost the opposite conditions prevail during the NAO-phase.
– The seesaw between the two NAO phases influences the northern Eurasian and American climate in a different manner: In the NAO+ phase, northern Siberia is wetter and northern Canada is drier. This influences the runoff from the big rivers in both areas and modifies the freshwater balance of the polar basin, and thus the water exchange with the mid-latitudes (Peng and Mysak, 1993; Clark et al., 1999).
– Salinities along the two inflow branches (Barents Sea throughflow and West Spitzbergen Current) appear to have declined as the NAO evolved from its negative values in the 1960s to its highly positive ones in the 1990s, a fact which is consistent with the increasing freshwater accession, the increasing volume flux of sea-ice from the Arctic and the reduction in total sea-ice over this period (Dickson et al., 2000).

Finally, we have to point to the fact that some important features are possibly not adequately represented in Figures 9a and b. For instance, little is still known on how the NAO interacts with dynamical processes over northeastern South America. Additionally, other aspects like connections between the NAO and biological processes in a wide range of European terrestrial and freshwater systems could not been addressed here. More information about this topic can be found, for instance in Mann and Drinkwater (1994), Metz and Myers (1994), Fromentin and Planque (1996), Alheit and Hagen (1997) and Reid et al. (1998).

6. Temporal Structure of the NAO

If the seesaw between the Icelandic Low and the Azores High is a major factor for the variability of climate in Europe, then the idea of developing a measure for its state or strength is straightforward (Walker and Bliss, 1932). This NAO measure can serve as a simple diagnostic and it provides a climatological time series for further investigation. The same also applies for an index of the AO. As mentioned above, different indices were calculated based on suitable SLP data back in time (gridded data, station data at Azores and Iceland, and other station data). The type of the index also depends on the type of analysis: for comparison with climate models, for instance, a NAOI based on gridded data is preferred. Another reason is that different statistical concepts are involved. This chapter gives an overview on the various measures for the NAO (and for some related patterns), commonly denoted North Atlantic Oscillation Indices (NAOIs) and their extrapolation into the past. Thereafter, the temporal evolution of the NAOI series and the spectral characteristics of the time series are presented.

6.1. Concepts and definitions to measure the NAO

In chapter 3.1, we have given a short overview about the various concepts for defining an index for the NAO. In chapter 3.2, we have also noted that, in some cases, the various indices can differ quite substantially. If all of these indices were meant to measure the same thing, they do a bad job, at least in the case of August 1976 (Figure 7). Why is this, and how different are the indices? Here we would like to compare and correlate some well-known indices, partly following Wallace (2000). We use the station-based NAO indices by Rogers (1984), based on Ponta Delgada and Reykjavik station data (denoted RO), the one by Hurrell (1995a), based on Stykkisholmur and Lisbon station data (HU), and the one by Jones (Jones et al., 1997), based on station pressure data from Gibraltar and the Reykjavik area (JO). Furthermore, we include a gridded NAOI (65° N/20° W–60° N/15° W minus 40° N/30° W-35° N/25° W; denoted GR; Luterbacher et al., 1999; 2001a). From the zonal indices we use the index by Lorenz (1951; denoted LO), based on the zonal mean pressure gradient at 55° N. Three indices based on EOF analysis were included: the index by Walker and Bliss (1932; hereafter denoted WB), the first principal component of the SLP field poleward of 20° N (PC), and the AOI (Thompson and Wallace, 1998; denoted TW). An index based on the temperature difference between Norway and Greenland (denoted NG) is also used (Wallace, 2000).

One important problem consists in the time resolution of the indices. NAO indices can be defined for series of winter average pressure (November to March,

TABLE I

Matrix of correlations between various indices of the NAO-AO/NAM phenomenon, based on seasonal-mean December to March data for the period 1950 to 1994 (from Wallace, 2000). WB: Walker and Bliss index (1932); HU: Hurrell NAOI (1995a); NG: Temperature Norway Greenland (Wallace, 2000); PC: Subpolar SLP principal component; LO: Lorenz index (Lorenz, 1951).

	WB	HU	NG	PC	LO
WB	1	0.74	0.59	0.86	0.79
HU		1	0.37	0.83	0.87
NG			1	0.37	0.32
PC				1	0.95
LO					1

December to February), for monthly mean pressure, or for even shorter time periods such as 5 day or even one day means. Two station indices are perhaps sufficiently robust for defining seasonal means but can differ widely if used for shorter term sampling. However, by using SLP analyses over the whole North Atlantic it is possible to obtain robust NAO index on daily time scales (Stephenson et al., 2001).

A matrix of correlations of the winter (December to February) indices WB, HU, NG, PC, and LO for the time period 1950 to 1994 has been presented in Wallace (2000), along with the corresponding spatial patterns. The table is redrawn here as Table I. Not all correlations between these very different indices are high. WB is highly correlated to PC. This is not surprising since WB, as explained by Wallace (2000), is apart from a discretization step equivalent to an EOF. On the other hand, compared to an EOF based on gridded data, WB is biased by the station availability (not many Arctic stations). The index based on temperature differences shows a poor correlation to all other indices.

A similar table (Table II) was created which presents the correlation between various indices of the NAO-AO/NAM phenomenon (TW, GR, RO, JO, and HU; see definitions above) based on a winter (monthly-mean data from December to March) and in brackets an annual mean (monthly-mean data from January to December) for the period 1899 to 1997. As expected, the correlation between the 5 indices is highly positive and significant, both, during wintertime and for the entire year. The highest correlations were calculated between the standardised Rogers (1984) index (RO), the gridded NAOI (GR) and the Hurrell (1995a) index (HU), respectively. The lowest values were calculated between the AOI (Thompson

TABLE II

Matrix of correlations between various indices of the NAO-AO/NAM phenomenon based on monthly-mean December to March (in brackets monthly-mean January to December) data for the period 1899 to 1997. TW: AOI according to Thompson and Wallace (1998); GR: Gridded NAOI (65° N/20° W–60° N/15° W minus 40° N/30° W–35° N/25° W; Luterbacher et al., 2001a)). RO: normalized NAOI by Rogers (1984) based on the stations data of Ponta Delgada and Reykjavik; JO: NAOI by Jones et al. (1997) based on station pressure from Gibraltar and the Reykjavik area; HU: Hurrell (1995a) NAOI based on Stykkisholmur and Lisbon station data.

	TW	GR	RO	JO	HU
TW	1	0.69 (0.57)	0.76 (0.65)	0.78 (0.63)	0.77 (0.65)
GR		1	0.97 (0.95)	0.84 (0.74)	0.97 (0.94)
RO			1	0.85 (0.72)	0.98 (0.97)
JO				1	0.88 (0.77)
HU					1

and Wallace, 1998) and the gridded NAOI (TW and GR). Station based indices have the disadvantage that they do not capture well the NAO pattern when shifted longitudinally or latitudinally. However, this is exactly what occurs in the course of a year, since the centres of both the Icelandic Low and the Azores High move (see Wanner et al., 1997; Portis et al., 2001). Barnston and Livezey (1987) also show that the SLP pattern most associated with the NAO appears quite different during different seasons (see their Figure 2) with seasonally specific prototypes that define quite different centres of action, particularly during the transitional seasons. Their result actually undermines the utility of using a two stations based index with fixed locations, since the lobes of the NAO actually wander east-west and north-south from season-to-season. As mentioned above, Barnston and Livezey (1987) thus, wisely, favour a definition of the underlying atmospheric circulation patterns based on an orthogonal PCA rotation approach, rather than a teleconnection index approach.

6.2. Temporal structure

In this section, the temporal structure of the NAO is discussed using the instrumental data based index by Hurrell (1995a). In recent years, this index was the most often used one and it is widely accepted. Figure 10 shows the mean winter (DJFM) index for the period 1864–1998. The smoothed line denotes the low-pass filtered (weights 1,3,5,6,5,3,1) series. Fluctuations with periods less than 4 years are removed. Hurrell showed with this time series that the NAO often persisted in one phase over many winters. He further noted that the NAO index exhibited

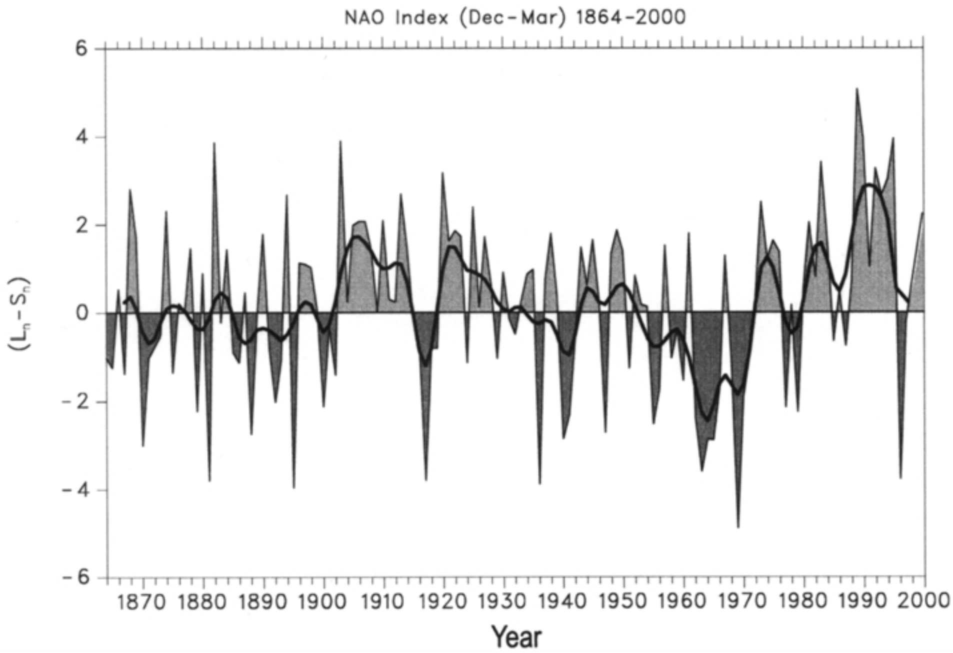

Figure 10. NAO index from 1864 to 2000, defined as the difference in normalized pressure between Lisbon and Stykkisholmur, for the winter months December to March (Hurrell, 1995a). Thick lines are 7-point low-pass filtered time series.

strong decadal trends. From the 1940s until the early 1970s, a strong downward trend was observed, whereas a sharp reversal towards strong positive values has occurred in the last 25 years. According to Hurrell (1995a), these strong positive index values from the mid 1980s to the mid 1990s were unprecedented in this index time series since 1864. He also emphasised that these strong events contributed significantly to the winter warming during these years across Europe. The atmospheric moisture budget in Europe was also greatly affected, leading to dry conditions over southern Europe and the Mediterranean and wetter than normal conditions in parts of Scandinavia and northern Europe. Based on homogenised surface pressure data, Slonosky and Yiou (2001) calculated new two-point NAO indices for winter (Gibraltar minus Reykjavik) and summer (Ponta Delgada minus Reykjavik) between 1820 and 2000. They were able to show that, by using this more precise data set, that the positive NAOI trend in recent years is not unprecedented in the twentieth century because similar positive values were also observed in the decades between 1900 and 1930 (see also Luterbacher et al., 2001a).

Figure 11 displays the histograms of the monthly NAOI by Hurrell (1995a), Jones et al. (1997) and Luterbacher et al. (2001a). By definition, the values of the index are centred around zero. The distributions are unimodal with a mode of zero. There were apparent periods with the NAO index being more or less in one phase.

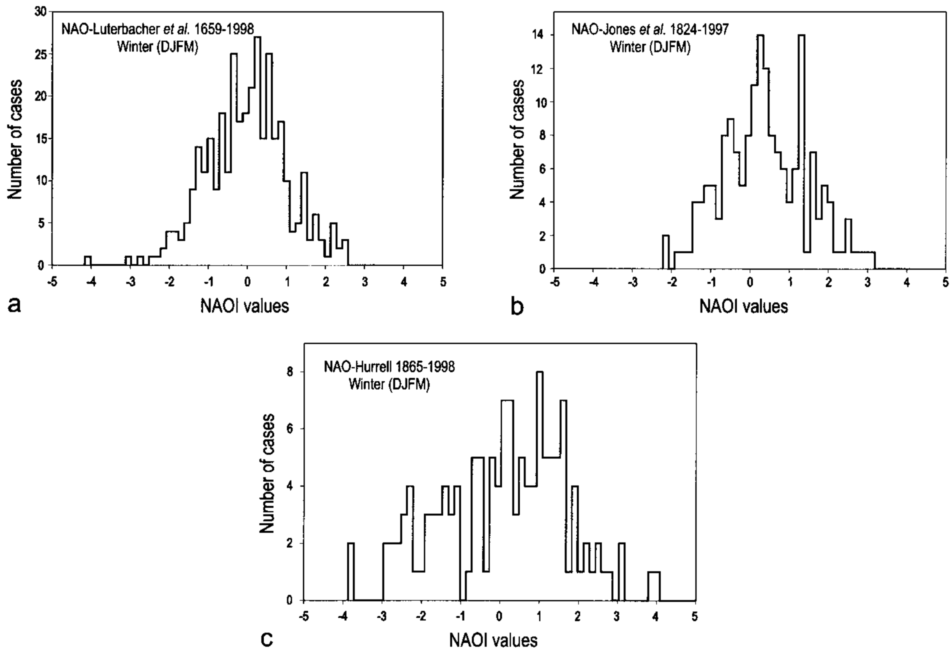

Figure 11. Histograms with the distribution of the monthly-mean (December to March) NAO indices by (a) Hurrell (1995a) from 1865–1998, (b) the Jones et al. (1997) NAOI from 1824 to 1997, and (c) the gridded NAOI of Luterbacher et al. (2001a) from 1659–1998. Note the different scaling on the y-axes.

Hurrell and van Loon (1997) found considerable variations of the NAO index at quasi-biennial and quasi-decadal timescales for the last 130 years.

Stephenson et al. (2000) showed with stochastic models that there is no contradiction between the noisy interannual behaviour of the NAO index and the observed long-range persistence. Simple stochastic models are capable of generating long-range processes with trends similar to those seen in the NAO. Stephenson et al. (2000) suggested that care should be exercised when assessing and attributing causes to trends in the NAO that may be due to simply natural long-range dependencies.

6.2.1. *Historical NAO Reconstructions*

Since it is not possible to calculate an instrumental NAO index much further back than 1821 with the direct use of pressure observations in the vicinity of Iceland and the Azores, different authors conducted (mainly winter) NAO index reconstructions for the past few centuries. Most of these reconstructions were based on tree-ring information from Europe, North Africa and North America (Cook et al., 1998; Cook et al., 2001; Glueck and Stockton, 2001) or based on ice core data from Greenland (e.g., Barlow et al., 1993; White et al., 1996; Appenzeller et al., 1998). Cullen et al. (2000) used tree-rings, ice cores and instrumental records for their multiproxy approach for 1750-1979. Luterbacher et al. (1999; 2001a, b) used observational data from the early instrumental period in combination with documentary proxy data for monthly reconstructions back to AD 1659 and seasonal reconstructions from AD 1500 to AD 1658. Rodrigo et al. (2001) reconstructed the winter Jones et al. (1997) NAO index from 1501 to 1997 based on reconstructed seasonal precipitation in Andalusia (southern Spain) from a wide variety of documentary data. Documentary sources for the Spanish journeys to America sailing times were used as proxies for reconstructing the NAO for the periods AD 1551 to 1650 and 1717 to 1737 (Garcia et al., 2000). The aim behind these reconstructions was to have a database to study the low frequency (predominantly natural) climate variability. Some studies dealt with rather long considerations or reconstructions related to the NAO phenomenon. Proctor et al. (2000) provided an annual NAO for the last 1100 years based on banded stalagmite growth rate from a cave in NW Scotland. Tremblay et al. (1997) used a dynamic-thermodynamic sea-ice model to simulate the modes of the Arctic Ocean ice circulation. Based on driftwood records they were able to suggest that for centuries to millennia during the Holocene, the high latitude average atmospheric circulation may have resembled that of 1968 (low NAO, weak Beaufort Gyre, broad Transpolar Drift Stream (TDS) and therefore large ice export), 1984 (high NAO, expanded Beaufort Gyre, weak TDS and low ice export) and today's climatology with abrupt changes from one state to the other.

Schmutz et al. (2000) compared different NAO proxy index reconstructions. Although all reconstructions were originally validated and thought to have a known reliability, they showed that most of the indices were poorly reconstructing the NAO variability if compared to an observational NAO index (e.g., Jones, 1997) in a temporally independent period in the first half of the nineteenth century. The most reliable reconstruction was the one by Luterbacher et al. (1999). In a second assessment, Schmutz et al. (2000) showed that the different winter NAO index reconstructions were mutually uncorrelated in the period 1716–1815. Their findings touch the serious question about the choice of significant calibration periods in a system with non-stationary forcing conditions.

Figures 12a and b display the seasonally (winter [DJFM]) and yearly [January to December] NAO index reconstruction by Luterbacher et al. (2001a). In general, the time series indicate clear evidence for the intermittent behaviour of the climate system in the North Atlantic region (Appenzeller et al., 1998) with a strong interannual variability. The NAO winter index (Figure 12a) shows low values at the end of the eighteenth and nineteenth centuries and between 1950 and the beginning of the 1970s. During the eighteenth century, the NAOI reveals clear decadal to interdecadal variations. Stronger westerlies (positive values) were prevalent from 1830 to 1870 and in the early twentieth century. The winter NAOI of the instrumentally based data of Hurrell (1995a; Figure 10) and Jones et al. (1997) reveals extremely positive values with three remarkable peaks between 1970 and 1995. These peaks are also visible in the data set of the Luterbacher et al. (2001a) in Figure 12a.

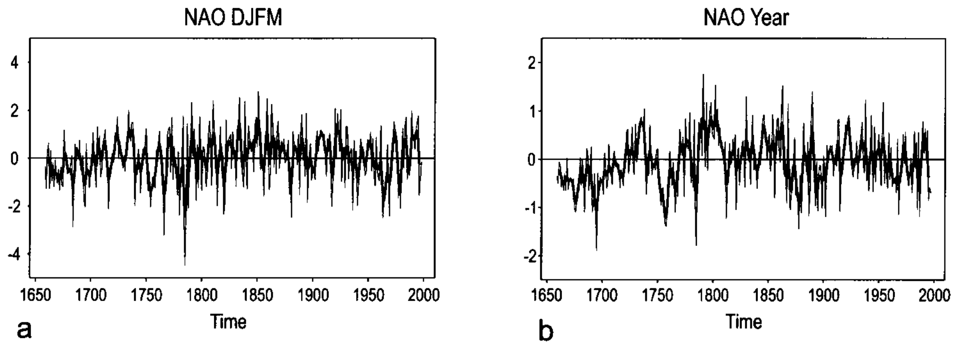

Figure 12. Normalised time series of the reconstructed: (a) winter (DJFM) and (b) annually (January to December) NAOI from 1659–1997 by Luterbacher et al. (2001a). The most recent seven values have been added from the Hurrell (1995) index (the correlation between the winter Hurrell (1995a) NAOI and the gridded winter NAOI is 0.97, see Table II). The thick line is the 7-point low-pass filtered series.

The annual values (Figure 12b) show a series of low indices at the end of the seventeenth century, after 1750, and between 1850 and 1900. The first period marks the end of the cold Maunder Minimum (Wanner et al., 1995). The third period between 1850 and 1900 is often considered to mark the end of the so-called 'Little Ice Age' in the North Atlantic.

6.2.2. *Spectral Analysis of NAOI Time Series*

Spectral decomposition of climatological time series can be used to study predictability: if clear periodicities exist, they can then be extrapolated into the future. Spectral analysis can also give insight into the dynamics of the climate system.

Several papers (e.g., Rogers, 1984; Hurrell and van Loon, 1997; Cook et al., 1998) touched on the subject of the spectral behaviour of the NAOI. Four papers (Appenzeller et al., 1998; Higuchi et al., 1999; Wunsch, 1999; Stephenson et al., 2000) concentrated on statistical considerations, mainly on the discussion of the temporal variability of the NAOI. Appenzeller et al. (1998) pointed to the intermittent character of the NAO with temporally active (coherent) and passive (incoherent) phases. By applying a Multiresolution Fourier Transform (MFT), Higuchi et al. (1999) analysed the temporal NAO variability. They found that the NAO displays fluctuations on multiple timescales and that the relative contributions of these components are not constant in time. The timescales range from interannual to interdecadal. In a recent study, Pozo-Vázquez et al. (2000) analysed the NAO variability since the beginning of the nineteenth century. They showed that, for winter, Gibraltar represents the southern pole of the NAO dipole better than do the Azores or Lisbon. By using cross-spectral analysis, they also showed that the most coherent out-of-phase variations between the two stations occur in the periods of 2.5, 5–6 and 8 years.

In order to test the predictability of the NAO, Stephenson et al. (2000) applied different statistical methods to show that the year-to-year differences in global land/sea surface temperature covariances are dominated by the NAO and to a lesser extent by the QBO. Wunsch (1999) emphasised that trends could be the result of a short sampling of noisy time series or due to natural or man-made non-stationarities in these time series. He showed that the NAO is weakly "red", with only slight broadband features near 8 and 2.5 years. Both, Wunsch (1999) and Stephenson et al. (2000) estimate that on average only about 10% of interannual variance in wintertime mean NAO can be forecasted one year ahead.

The same spectral features were already found by Baur (1927) in the 100 year temperature record of Berlin and were well established at that time. Recently, Eshel (2000) suggested that some NAO variability is externally forced from the Pacific. By using this fact, he demonstrated that at a 15 months lead, NAO forecasts are robust and skilful under a stringent cross-validation.

The spectral behaviour of the extended and updated indices reconstructed by Luterbacher et al. (2001a) described above is investigated here. First, the period 1659–1997 was subject to a wavelet analysis (Figures 13a, b). In order to test whether the reconstructed index time series show significant periodic oscillations or not, a global and a local wavelet spectral analysis with a Morlet wavelet base (Torrence and Compo, 1998) were performed. The statistical significance of a given local or global wavelet power spectrum was tested, with a 10% significance level, against the null hypothesis that the respective index stems from a white-noise process. For the winter and annual time series, the analysis was done for the period 1659–1997.

The local wavelet power spectra again show the typical intermittent behaviour of the NAO, with active phases and corresponding maximum amplitudes in different frequency bands. The winter spectrum (Figure 13a) does not clearly reveal the classical peaks at about 2.5 and 6–10 years as found by Hurrell and van Loon (1997) in their classical study for a limited period from the late nineteenth to the end of the twentieth century. It does at least represent the enhanced power at the period of 6 to 10 years between about 1975 and the beginning of the 1990s. Finally, on the annual base (Figure 13b), the spectrum reveals high power at frequency bands around 2-10 and especially 64 years. The 64 years peak has to be questioned even though it is also visible in Alpine precipitation series (Wanner et al., 2000) because at least 10 cycles are normally required to get good estimates. At least, it is noteworthy that Delworth et al. (1997), Schlesinger and Ramankutty (1994) as well as Delworth and Mann (2000), based on both, modelling and observational studies, found very similar periodicities.

Similar to Wanner et al. (1997), Portis et al. (2001) studied the intraseasonal mobility of the NAO dipole. They defined a seasonally and geographically varying "mobile" NAO index and showed that these NAO nodes maintain their correlation from winter to summer better than traditional NAO indices. The question is how

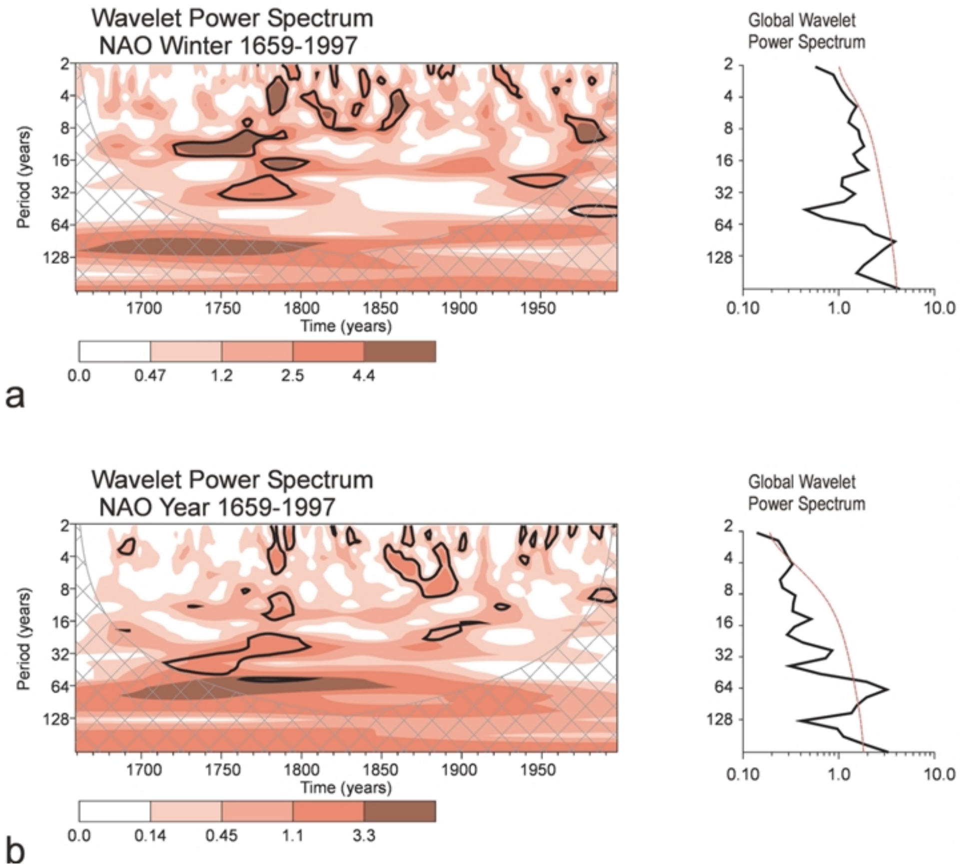

Figure 13. Wavelet curves (annual mean (January to December) and winter mean (DJFM)) of the monthly Luterbacher et al. index (2001a) from 1659 to 1997. The last seven values stem from the Hurrell NAOI (1995a). Amplitudes are scaled with variance of the index. The logarithmic vertical axis indicates equivalent periods; the horizontal axis represents time. The 90% confidence limits (based on the global wavelet on the right) are given in thick contour lines.

long the phenomenon can be called NAO. Multivariate NAO indices over the whole of the North Atlantic are also less sensitive to such shifts (Stephenson et al., 2001).

It is quite speculative to infer that a specific dynamical process was the cause of a certain quasi-periodicity. We would merely like to emphasise that different authors pointed to possible mechanisms representing specific timescales and occurring in atmosphere, ocean or in the coupled atmosphere–ocean–sea-ice system (see also Section 4). For the 2.5-year quasi-periodicity, one may speculate that this might be a consequence of the QBO (Reed et al., 1961). For the period of 6 to 10 years, oscillations in the Arctic ocean–sea-ice–atmosphere system could play an important role (Mysak and Venegas, 1998). In the case of the 32 and 64 years periodicities, in a first paper, Cook et al. (1998) reported about a 70 year ocean induced variability which was no more clearly detectable in a second paper (Cook et al., 2001). Stocker (1996) mentions different mechanisms (e.g., the GC-THC dynamics) being responsible for the generation of processes which can be assigned to this timescale. Schlesinger and Ramankutty (1994) attributed the probable cause of a 60 to 70-year periodicity to a possible internal oscillation of the atmosphere–ocean system. This is in agreement with Enfield et al. (2001), who report about a 65 to 80 year oscillation of the North Atlantic SSTs, which is believed to have had a large impact on European and American climate in the nineteenth and twentieth centuries and may be driven by the Atlantic THC. Delworth et al. (1993; 1997) showed that the THC has an irregular oscillation with a characteristic timescale of approximately 40 to 60 years by involving large-scale interactions between Arctic freshwater and ice export, and Higuchi et al. (1999) suggested that the North Atlantic SSTs possibly modulate the NAO at a timescale of about 60 years. Just recently, Mokhov et al. (2000), by interpreting two 1000-year numerical simulations and the data by Jones et al. (1997) and Luterbacher et al. (1999), stated that the regimes of quasi-decadal variability of the NAO differs clearly in different centuries.

What could we learn from the temporal behaviour of the NAOI? The NAO is not a pure random process, nor does it exhibit a clear mode of oscillation. The autocorrelation in the index time series and the weakly significant and intermittent quasi-biennial and quasi-decadal oscillations of the NAO index are strong arguments for a low frequency impact of the North Atlantic ocean on the atmosphere. It is finally interesting to note that the NAO can be characterised as having a weakly red spectrum with some long-range dependence.

7. Simulated Future NAO Trends

Atmospheric and climate models have been widely used to study aspects of the Atlantic-European climate, in particular atmospheric variability and ocean-atmosphere interactions (e.g., Palmer and Sun, 1985; Frankignoul, 1985; Lau and Nath, 1990; Ferranti et al., 1994; Lau and Nath, 1994; Peng et al., 1995; Kushnir and Held, 1996; von Storch et al., 1997; Broccoli et al., 1998; Latif, 1998; Itoh and Kimoto, 1999). Here the focus was mainly on the capability of GCMs and AOGCMs to simulate possible future changes of the NAO or AO index.

Most published analyses of climate change simulations give information on the NAO for winter only. The only exception known to us is the study by Liang et al. (1996) who found a northwest shift of the entire NAO-pattern for April in a 2 $\times CO_2$ simulation with the National Center for Atmospheric Research/Community Climate Model (NCAR/CCM1-GCM). In four different AOGCM simulations the winter NAO index was found to exhibit either no (Fyfe et al., 1999), a slightly positive (Ulbrich and Christoph, 1999), or a slightly negative (Osborn et al., 1999) trend. In two further simulations, Campbell et al. (1995) detected a western shift of the southerly centre of action (Azores High) of the annual mean (showing a strong bias to the winter) NAO-pattern. Three further studies showed a general weakening (Liang et al., 1996), respectively southern (Huth, 1997) and northeastward (Ulbrich and Christoph, 1999) displacement of the entire winter pattern. It is noteworthy, that in most simulations the projected NAO behaviour was accompanied by distinct changes in the atmospheric circulation. For instance, Fyfe et al. (1999) detected a positive trend in the AO-index, Liang et al. (1996) found a simultaneous weakening of the Siberian activity centre of the EU1, one of the two Eurasian patterns defined by Barnston and Livezey (1987). Huth (1997) detected shifts in both the EU1 and EU2 patterns of Barnston and Livezey (1987) pattern, and Ulbrich and Christoph (1999) found increasing upper air storm track activity over the east Atlantic and western Europe with rising GHG forcing. Gyalistras (2000) analysed 33 studies presenting climate change simulations of 24 models, published by researchers of 7 research centres during the last 15 years. For a CO_2 doubling scenario (without consideration of sulphate aerosols) 4 model runs revealed a decreasing, 2 a similar and 8 an increasing intensity of the storm tracks in the Atlantic-European area.

By simulating recent greenhouse-gas forced northern winter climate trends, Shindell et al. (1999) found that the AO is captured only in climate models that include a realistic stratosphere, while changes in the ozone concentrations are not necessary to simulate the observed trends. In general, they showed that the observed trend since the 1970s with a gradual reduction of high-latitude sea-level pressure, and with an increase of mid-latitude sea-level pressure and AO may continue. Based on observations and model data sets with different forcing mechanisms, Paeth et al. (1999) demonstrated that the increasing CO_2 concentration has a significant influence on the simulated NAO variability on time scales of 60 yr and longer. On time scales less than 10 yr, the interannual variability of the NAO states decreases synchronously with the positive trends of its decadal-mean state implying a stabilisation of its pressure and future zonal state.

If one assumes that coupled modes play an important role on the decadal time scale, the question about the future trends of ocean meridional heat transport related to THC has to be posed. Stocker et al. (2001) emphasise that paleoclimate models show that future changes of the THC in response to global warming are likely. The magnitude of the change is highly uncertain, but the models agree that the THC in the Atlantic will reduce due to the gain of buoyancy associated with the warming and a stronger hydrological cycle.

8. Summary and Conclusions

The NAO, a large-scale meridional displacement of atmospheric mass between the North Atlantic regions of the subtropical anticyclone near the Azores and the subpolar low pressure system near Iceland, is a major source of seasonal to interdecadal climate variability in the Northern Atlantic and European regions (Exner, 1924; Walker, 1924; Rogers, 1984; Lamb and Peppler, 1987; Hurrell, 1995a). It is often defined as the difference between the normalised mean winter (December to March) SLP anomalies between a station in or around the Azores and Iceland. The NAO is well correlated with the AO/NAM, which is defined as the leading EOF of monthly SLP fields north of 20° N weighted by area (Thompson and Wallace, 1998). The AO/NAM is zonally more symmetric than the NAO, but nevertheless shows a clear predominance in the Atlantic-European region (Deser, 2000). In the view of the NAO, the pressure distribution over the Atlantic region and its connection with the ocean and sea ice dynamics is central. The AO/NAM view assumes that the main actor is the Northern Hemispheric zonal circulation with its Arctic centre of action (Wallace, 2000). In the historical view, the NAO and AO/NAM were not developed separately from each other. In our opinion, both concepts have a legitimacy to exist. If the AO/NAM concept helps us to consider the planetary-scale processes in the upper troposphere and stratosphere, whereas the NAO concept has stimulated attention on North Atlantic ocean–atmosphere–sea-ice interaction (Sections 1 and 2).

Periods with predominating mild or severe winters in Europe caught the attention of westward moving northern Europeans. In the nineteenth century, several European climatologists started to study the underlying processes, namely the SLP and temperature seesaw (Dove, 1839, 1841; Hann, 1890; Teisserenc de Bort, 1883). Petterson (1890) and Meinardus (1898) first investigated the influence of the interannual fluctuations of the Gulf Stream on climate and weather in Western Iceland and Greenland. Together with Hildebrandsson (1897), who studied the inverse relation between the Azores and Iceland SLP, they were the predecessors of pioneers like Exner, Walker, Rossby, Lorenz and Bjerknes (Section 3.). The temperature seesaw between Greenland and northern Europe has been noted already in the eighteenth century. The seesaw was verified during the nineteenth century with meteorological time series and was based on physical principle: the strength and position of the centres of action. While oceanographers started to study the dynamics of the Gulf Stream and to speculate about its influence on climate variability over Europe, others started to study the pressure variations of the entire NH using descriptive statistical techniques.

The stationary wave in the lee of the Rockies and the North American continent that is set up by the orography, by diabatic heating, and by transient forcing, forms an important dynamical background for the statistical NAO phenomenon. It should be asked what is the dynamical influence of the Greenland land mass and the Arctic Ocean basin. A large number of modern studies investigate additional atmosphere-alone or coupled atmosphere–ocean–sea-ice processes that can produce NAO-like fluctuations.

As already mentioned, several authors have asked the question whether the NAO is a purely atmospheric phenomenon or not. Based on modelling studies, James and James (1989) showed that the atmosphere alone is able to create low-frequency variability, and Barnett (1985), together with Marshall et al. (1997) demonstrated that it is possible to reproduce NAO-like atmospheric variability with temporally non varying SSTs. Even though different mechanisms are postulated to be responsible for the creation of purely atmospheric NAO, we are still far from a consensus. Longer-term variability can be traced back to troposphere-stratosphere coupling (Perlwitz and Graf, 1995; Kodera and Kuroda, 2000; Perlwitz et al., 2000), but what is the trigger behind the complex troposphere-stratosphere interaction (QBO, solar or volcanic forcing, ozone depletion, greenhouse effect; see Robock and Mao, 1992; Labitzke and van Loon, 1995; Graf et al., 1998; Baldwin and Dunkerton, 1999; as well as Section 4)?

A series of observational studies important for the diagnosis of the NAO/AO-NAM phenomenon deals with one way atmosphere-ocean (Hasselmann, 1976; Wallace et al., 1990; Zorita et al., 1992; Frankignoul et al., 1997) or ocean-atmosphere forcings (Ratcliffe and Murray, 1970; Palmer and Sun, 1985; Peng et al., 1995; Czaja and Frankignoul, 1999; Rodwell et al., 1999; Bretherton and Battisti, 2000). On interannual time scales, atmospheric (wind stress) anomalies might be able to force the variations in the ocean by anomalous heat fluxes and surface mixing. The communication of the upper ocean anomalies like the SSTs to the atmosphere through evaporation, precipitation and heating processes might be a possible factor influencing the longer-term NAO dynamics. It seems that only pronounced positive (negative) SST anomalies can reproduce a barotropic atmospheric response with positive (negative) upper level pressure anomalies in the lee of the warm (cool) pool in the ocean.

Because most of the observational data series are too short to provide a sound basis for decadal or interdecadal studies dealing with coupled ocean–sea-ice–atmosphere modes, such studies are mainly centred on modelling and model validation (Latif, 1998). Despite many inconsistencies in the different approaches, there is some evidence that interdecadal Atlantic climate variability can be simulated if the main dynamics of the ocean circulation systems (THC, subtropical to subpolar GC) are included in fully coupled ocean–sea-ice–atmosphere models. Delworth et al. (1993; 1997) showed that the THC has an irregular oscillation with a timescale of approximately 40–60 years, which is driven by density anomalies in the sinking region of the THC. These SST anomalies induce surface-air temperature anomalies over the North Atlantic, the Arctic and northwestern Europe. Timmermann et al. (1998) demonstrated that an anomalously strong THC can produce positive SST anomalies and strengthen the NAO which itself reinforces the freshwater fluxes in the subpolar GC, and therefore induces a damping of the THC (cycle length: 35 years). Grötzner and Latif (1998) showed that an interaction between the wind-driven GC and the atmospheric circulation might also cause decadal climate fluctuations in the North Atlantic. There is no doubt that interactions between the midlatitude Atlantic Ocean and the tropical Atlantic Ocean (Chang et al., 1997), as well as between the North Atlantic and the Arctic Ocean (Dickson et al., 2000; Venegas and Mysak, 2000) play an important role for the understanding of the NAO/AO-NAM variability. Figure 9 shows the attempt to combine important knowledge about the NAO/AO-NAM phenomenon graphically (see also the Sections 4 and 5).

Was the NAO mode temporally stable in the past and how does it react in the future? Based on an index determined with measured SLP time series from 1864 through 1994 Hurrell (1995) showed that the NAO often persisted in its positive or negative phase over several winters and exhibits clear decadal trends. Hurrell and van Loon (1997) found considerable NAOI variations at quasi-biennial and quasi-decadal timescales for the last 130 years. From the 1940s to the 1970s, a strong downward trend was observed, and a sharp reversal occurred in the last 25 years. Slonosky et al. (2001) demonstrated that the positive NAO trend in recent years is not unprecedented in the twentieth century because similar positive values were also observed between 1900 and 1930.

Luterbacher et al. (1999; see also Mokhov et al., 2000) studied a longer time series from 1675 to 1997. The local wavelet spectra showed the typical intermittent behaviour of the NAO. The winter spectrum does not clearly reveal the classical peaks at about 2.5 and 6–10 years. Anyhow, a slightly significant power is visible at the period of 6 to 10 years between 1975 and the 1990s. On the annual base, the spectrum shows high power at frequency bands around 2–10 and especially 64 years (Section 6).

Most simulations of the future NAO only give information for winter. Based on the existing studies, no clear statement can be made for a CO_2 doubling because some models reveal a strengthening and northward displacement (positive NAO), and some others a southward movement of the westerlies (negative NAO). Gyalistras (2000) reported that more models simulating a CO_2 doubling reveal an increasing frequency with a northward movement of the storm track over the Atlantic Ocean and Europe (Section 7).

In conclusion, a clear indication is given that – especially for the understanding of decadal to century scale NAO/AO-NAM variability – more significant studies based on longer-term proxy based reconstructions, model runs with coupled AO-GCMs, and studies on underlying high- to low-frequency mechanisms are needed in the future.

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