The North Atlantic Oscillation

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Abstract.

The North Atlantic Oscillation (NAO) is the most important mode of variability in the northern hemisphere (NH) atmospheric circulation. Put simply, the NAO measures the strength of the westerly winds blowing across the North Atlantic Ocean between $40^o N$ and $60^o N$. The NAO is not a regional, North Atlantic phenomenon, however, but rather is hemispheric in extent. Based on 60 years of data from 1935 to 1995, Hurrell(1996) estimates that the NAO accounts for 31% of the variance in hemispheric winter surface air temperature north of 20^oN. The present article provides an overview of the NAO, its role in the atmospheric circulation, its close relationship to the Arctic Oscillation of Thompson and Wallace(1998), and its influence on the underlying North Atlantic Ocean. Some discussion is also given on the dynamics of the NAO, the possible role of ocean surface temperature, and recent evidence that the stratosphere plays an important role in modulating the NAO.

1. Introduction

The North Atlantic Oscillation (NAO) is the most important mode of atmospheric variability over the North Atlantic Ocean, and plays a major role in weather and climate variations over Eastern North America, the North Atlantic and the Eurasian continent (van Loon and Rogers, 1978; Wallace and Gutzler, 1981; Hurrell, 1995b, 1996; Kushnir, 1999). The NAO was first identified in the 1920's by Sir Gilbert Walker (Walker,1924; Walker and Bliss, 1932). Most attention has been focused on the winter season. It is, nevertheless, an important feature of atmospheric variability throughout the year (Barnston and Livesey, 1987), although it is less dominant during the warmer seasons (Rogers, 1990). Put simply, the NAO is a measure of the strength of the westerly winds blowing across the North Atlantic Ocean in the $40^oN - 60^oN$ latitude belt (see Figures 2 and 7). During winters when the NAO index is high, the westerly winds are stronger than normal. The moderating influence of the North Atlantic Ocean then leads to warmer than normal conditions over the Eurasian continent, while the eastern Canadian Arctic is colder than normal (Hurrell,1995b,1996). Rogers(1984) notes that the NAO is nicely summarized by Walker(1924): "... it is generally recognised that an accentuated pressure difference between the Azores and Iceland in autumn and winter is associated with a strong Gulf Stream, high temperatures in winter and spring in Scandinavia (Meinardus, 1898) and the east coast of the United States, and with lower temperatures in the east coast of Canada and the west of Greenland". Although Walker's assertion about the Gulf Stream may not be correct, the seesaw in mean winter temperature between Greenland and northern Europe is a robust feature of the NAO that has been known since the 18th Century (see van Loon and Rogers, 1978, and Loewe, 1937, 1966). Indeed, van Loon and Rogers(1978) and Rogers and van Loon(1979) take the seesaw as their starting point by classifying winters according to the winter temperature anomalies in Jakobshavn, Greenland, and Oslo, Norway. High NAO index winters are also associated with drier conditions over much of central and southern Europe, and wetter than normal conditions over Iceland and Scandinavia. Understanding the NAO and its variability is therefore of considerable socio-economic importance.

The role of the ocean, and in particular sea surface temperature (SST), in regulating the NAO has attracted much attention (see, for example, Kushnir, 1999), but remains controversial. An important aspect of the NAO is that dynamical coupling with the ocean is not an essential feature of its dynamics. In fact, interannual variability of the NAO is found in atmospheric general circulation models (AGCM's) which are run with specified, seasonally varying SST (Barnett, 1985). The typical limit of predictability for the atmosphere is about three weeks. Three weeks is, therefore, the likely intrinsic limit of predictability for the NAO (although see comments at the end of Section 6 regarding quasistationary regimes of the atmospheric circulation). On the other hand, the large heat capacity of the ocean compared to the atmosphere gives the ocean a much longer memory of its past state. It follows that if the variability of the NAO is somehow driven by that of the underlying SST, then there is hope for predictive capability on longer time scales, perhaps out to seasonal or even beyond (e.g. Griffies and Bryan, 1996). An important caveat is that the SST must itself be predictable, an issue we address in Section 4 in connection with recent work of Bretherton and Battisti(2000).

In this article, an overview is given of the NAO, its role within the atmospheric circulation (Section 2), its role in driving the ocean (Section 3), its likely dynamics (Section 4), including the possible role of SST, and in Section 5 a discussion of the "Null Hypothesis" for North Atlantic climate variability. Section 6 provides a brief summary and conclusions.

2. The NAO and the atmospheric circulation

Hurrell(1995b, 1996) has defined an index for the NAO as the difference between normalised mean winter (December to March) sea level pressure (SLP) anomalies at Lisbon, Portugal and Stykkisholmur, Iceland (Hurrell, 1995b). The normalisation is achieved by dividing the SLP anomalies at each station by the long term (1864-1994) standard deviation. Other, essentially equivalent, indices of the NAO have also been constructed. For example, Rogers(1984) uses the difference between normalised SLP anomalies at Ponta Delgadas, Azores, and Akureyri, Iceland, and defines winter as December to February, excluding March. There have also been attempts to extend the index back beyond the instrumental record using tree-ring and ice-core data (e.g. Appenzeller et al., 1998; Cook et al., 1998; Cullen et al., 2000; Luterbacher et al., 1999; Stockton and Glueck, 1999).

Figure 1 shows the time series of Hurrell's NAO index from 1864-1994, and Figure 2 shows the winter SLP departure associated with one standard deviation (positive) of the NAO index. Geostrophic balance implies that when the index is high, the westerly winds across the North Atlantic are stronger than normal, whereas when the index is low, the westerly winds are weaker than normal.

The observed surface temperature (SST and air temperature over land) departure associated with one standard deviation (positive) of the NAO index is shown in Figure 3. The seesaw in winter temperatures between western Greenland and Europe is evident, a high NAO index (stronger than normal westerly winds) being associated with cold winters in western Greenland and warm winters in Europe, and vica versa (van Loon and Rogers, 1978). The importance of the NAO for understanding winter temperature variability is clear from the analysis of Hurrell(1996). He has shown that the NAO alone can account for 31% of the winter surface temperature variance over the northern hemisphere north of 20^oN, and that the NAO and El Nino combined can account for 44% of the winter surface temperature variance (based on 60 years of data from 1935 to 1994). The NAO is therefore important not just for winter surface temperature variability in the North Atlantic sector, but for winter surface temperature variability over the northern hemisphere as a whole. Indeed, it is a remarkable feature of Figure 3 that the change in winter temperature associated with the NAO extends all the way across the Eurasian continent from the Atlantic to the Pacific, with the NAO apparently having just as much influence on winter temperature in Siberia as in western Europe. This is an indication that the NAO is not a regional North Atlantic phenomenon, but is actually closely related to a hemispheric mode of variability known as the Arctic Oscillation (Thompson and Wallace, 1998).

As defined by Thompson and Wallace(1998), the Arctic Oscillation (AO) corresponds to the first EOF (Empirical Orthogonal Function) of SLP variability over the northern hemisphere. As can be seen from Figure 4 (bottom panel), the spatial structure of the AO corresponds closely to that associated with the NAO (Figure 2). Thompson and Wallace(1998) find that the correlation coefficient between the principal component time series (the AO index) and the NAO index is 0.69, and show that the AO index is actually more strongly coupled to Eurasian winter surface air temperature than the NAO index. It follows that while the AO and NAO are not identical, they are certainly closely related, and the AO index may, in fact, be a better indicator of the atmospheric mode of variability usually associated with the NAO. Accordingly, in what follows, we shall assume that the AO and NAO correspond to the same physical phenomenon, while bearing in mind the differences in detail. (For further discussion on the relationship between the AO and the NAO, and also on the nature of the AO itself, see Deser(2000) and Monahan et al.(2000).) Thompson and Wallace(1998) note that the AO accounts for 22% of the variance in winter SLP over the northern hemisphere poleward of 20^oN (here winter is defined as November-March, and 40 years of data are used from 1958 to 1997). They show that the AO is an equivalent barotropic phenomenon (see Figure 4), with a strong signature in the winter stratosphere where it accounts for 50% of the variance in the height of the 50-hPa pressure surface. Furthermore, the geopotential height anomalies at the 50-hPa level associated with the AO are almost five times as strong as the corresponding 1000-hPa height anomalies, and since the spatial patterns are almost the same, it follows that the energy density (the product of density and squared amplitude) is nearly the same at both 1000-hPa and 50-hPa. This remarkable feature of the AO has led some authors to suggest that the stratosphere may play an important role in its dynamics, as issue we shall return to in Section 4. The work of Thompson and Wallace(1998) also makes it clear that the variability in the AO consists of a transfer of mass in and out of the circumpolar polar vortex. A similar mode of variability (the Antarctic Oscillation or AAO) is known to exist in the Southern Hemisphere where the signature in the troposphere is even more symmetric about the pole (Thompson and Wallace, 2000). It seems likely that the bias towards the North Atlantic in the surface and middle troposphere structure of the AO is a consequence of the land-sea contrasts in the Northern Hemisphere (Thompson and Wallace, 1998) and presumably also the mountain ranges (e.g. The Rockies and Greenland) that cut across the path of the tropospheric jet stream.

Let us now examine the time series of the NAO index in more detail (Figure 1). Striking features are the low values during the period from the early 1950s to the early 1970s, relatively high values in the early part of this century, and the high values of the last 25 years, during which time the index also shows strong decadal variability. Caution should be exercised when interpreting these changes in the character of the index. Figure 5 shows the estimated power density spectrum computed from Hurrell's time series by Wunsch(1999). This shows a weakly red spectrum, with weak structures near periods of 2 and 8-10 years, but no particularly striking peaks. As such, the characterisation of the NAO as an "oscillation" is misleading since its spectrum is much more akin to that of white noise. Wunsch also shows that statistically there is no reason to attach any significance to the variations in the character of the time series, such as the periods of high and low index noted above, or the enhanced decadal variability in recent years. He shows that such events are the expected behaviour of a weakly red noise process, and need have no "cause" any more than does a sequence of rolls of dice that produces a statistical excess of the value six. This is an issue we shall return to in

Section 4.

The pattern of temperature changes in the Northern Hemisphere during the period 1981-1997 has been one of warming over the continents and cooling over the oceans (Figure 6), strongly resembling the pattern of temperature change associated with the NAO (Figure 3). The similarity between these two patterns suggests that the recent increase in temperature over the Northern Hemisphere is related to the positive phase of the NAO index since the early 1980s evident from Figure 1 (Hurrell, 1996; Thompson, Wallace and Hegerl, 2000). Some model simulations show an increasing trend in the AO index in response to greenhouse gas forcing (Shindell et al., 1999; Fyfe et al., 1999), leading to speculation that the recent tendency for a higher AO/NAO index might be associated with global warming. At the present time, however, this remains speculation (see, for example, Paeth et al., 1999).

The NAO also has an influence on storm track variability. Figure 7, taken from Rogers(1990), shows the January mean SLP and winter (December/January/February) storm tracks obtained by compositing months with high and low NAO index (see Rogers, 1990, for details). It is clear that the NAO has a major influence on storm tracks, with reduced activity in the northeastern part of the North Atlantic when the index is low.

The change in the mean and eddy components of the flow between high and low index also leads to a change in the convergence of moisture transport and is thus linked to regional precipitation variations. Figure 8 shows evaporation minus precipitation (E-P) computed as a residual of the atmospheric moisture budget using ECMWF analyses for high index minus low index winters (see Hurrell, 1995b, for details). The moisture convergence over the northern part of the British Isles and Scandinavia is associated with wetter than normal conditions in high index years, while high index is also associated with drier than normal conditions over parts of southern Europe, the Mediterranean and parts of North Africa (Hurrell and van Loon, 1997). The 1981-1994 winter average precipitation anomalies, expressed as departures from the

1951-1980 mean, show positive anomalies over the British Isles and northern Europe, and negative anomalies over central and southern Europe. This pattern bears a striking resemblance to the changes in precipitation corresponding to a unit deviation of the NAO index calculated for the winters of 1900-1994 (Hurrell and van Loon, 1997). As with temperature, this similarity suggests that the recent precipitation anomalies over Europe can be associated with to the upward trend of the NAO index during the past several decades. Thompson, Wallace and Hegerl(2000) show that many climate variables exhibit trends in the past several decades, and that a large part of the trend (although not all) can be accounted for by the upward trend in the AO. The changing precipitation patterns of the 1980's and 1990's led to reduced precipitation over the Greenland Ice Sheet and the Alps, but enhanced precipitation over the Scandinavian mountains with consequences for the local snow pack, and the associated skiing industry. The increased snowfall over the Alps in recent years can, in turn, be associated with a return to low NAO index conditions.

3. The NAO and variability in the North Atlantic Ocean

The Labrador Sea and the Greenland/Iceland/Norwegian (GIN) Seas of the North Atlantic Ocean are two of the few places where the deep waters of the world ocean are known to be renewed. The only other locations of any significance are the Arctic Ocean and the Ross and Weddell Seas around Antarctica. This means that unlike the North Pacific Ocean, the North Atlantic is an active conduit through which newly formed dense waters spread into the rest of the global ocean. The large scale flow associated with the spreading of newly formed dense waters is known as the thermohaline circulation (THC). Broecker(1991) has suggested that the THC associated with North Atlantic Deep Water forms a global ocean circulation he has termed the "Conveyor Belt". Associated with the "Conveyor Belt", heat is transported northward in both the North and South Atlantic Oceans. At 24^oN, the northward heat transport has been estimated as 1.2 ± 0.3 PW (Hall and Bryden, 1982) (1 PW = 10^{15} Watts). Almost all this heat is released to the atmosphere over the North Atlantic north of 24^oN, an amount of heat roughly equivalent to having one 100 Watt light bulb pointing upwards on every square meter of ocean! The heat released by the North Atlantic to the atmosphere plays an important role in moderating the climate of western Europe and, to some extent, Atlantic Canada. When the NAO index is high, the westerly winds penetrate more strongly into the Eurasian continent, and the moderating influence is correspondingly more effective than when the index is low (Figure 3).

Given the importance of the NAO/AO as a mode of variability in the overlying atmosphere, we should not be surprised that the NAO is an important forcing for the North Atlantic Ocean. This was recognised by Bjerknes(1964) in his now classic study of air-sea interaction over the North Atlantic. Indeed, Bjerknes made considerable use in his analysis of the surface pressure difference between Iceland and the Azores (an index of the NAO). Bjerknes examined the relationship between anomalies in North Atlantic SST and North Atlantic SLP. He suggested that on the interannual time scale, SST anomalies are locally driven by changes in the heat flux from the atmosphere, whereas on the decadal time scale, changes in the ocean circulation, and hence ocean heat transport play a role. The decadal time scale is the time scale associated with oceanic advection and baroclinic Rossby wave propagation and hence with the baroclinic adjustment of midlatitude gyres to changing surface forcing. Bjerknes' work has since been updated, and his conclusions supported by Cayan(1992), Deser and Blackmon(1993), Kushnir(1994), Battisti et al.(1995), Halliwell and Mayer(1996), Halliwell(1997, 1998), and by the modelling work of Häkkinen(1999), Eden and Jung(2000) and Eden and Willebrand(2000). Cayan shows a direct connection between North Atlantic SST (NASST) anomalies and forcing by the NAO. Bjerknes' suggestion is also consistent with results from the GFDL coupled ocean/atmosphere model. On the interannual time scale, NASST anomalies in the model are directly forced by surface flux variations from the atmospheric model (Delworth, 1996), whereas on the interdecadal time scale, NASST anomalies are associated with changes in the poleward heat transport associated with fluctuations in the THC of the model (Delworth et al., 1993). On both time scales, the important atmospheric forcing for the ocean model is closely related to the NAO in the atmospheric model (Delworth et al., 1993; Delworth, 1996; Delworth and Greatbatch, 2000).

Eden and Jung(2000) describe results from a North Atlantic circulation model driven by surface flux forcing based on a monthly NAO index from 1865-1997. These authors are the first to show conclusively that a circulation model has skill at reproducing the observed interdecadal evolution of SST anomalies in the North Atlantic, and to demonstrate the role played by the changing ocean circulation in that evolution. To obtain the spatial structure of the forcing for the model, they regress the fields of anomalous surface heat, freshwater and momentum fluxes against the NAO index for the period 1957-1997 using NCEP/NCAR reanalysis data (Kalnay et al., 1996). The spatial structure obtained from the regression is then multiplied by a monthly NAO index obtained as the difference in normalised monthly mean SLP between the Azores and Iceland (as in Rogers, 1984) for every month from 1865 to 1997. In this way, realistic forcing fields are obtained for the model from 1865 to 1997. The skill of the model at capturing the evolution of interdecadal SST anomalies is illustrated in Figure 9. Of particular interest is the cold anomaly that is found to the southeast of the US in 1960-64 and subsequently spreads and intensifies northwards in both the model and the observations. The spreading and intensification of this anomaly across 40^oN takes place in the model despite the absense of NAO-derived forcing near 40^oN, and, until the early 1980's, in the face of forcing of the opposite sign (i.e. anomalous heating of the ocean) north of 40^oN. In the model, the evolution of this SST anomaly is driven by the convergence of heat transport associated with the ocean circulation, and in particular by changes in the strength of the THC and the subpolar gyre circulation. In contrast to the interpretation given by Hansen and Bezdek(1996) and Sutton and Allen(1997), advection of the SST anomalies by the mean circulation is not important in this model (see also Visbeck et al., 1999). Eden and Jung also show that on the interdecadal time scale, the most important contribution to the anomalous forcing of their model comes from the surface heat flux, in agreement with the analysis of the GFDL coupled model carried out by Delworth and Greatbatch(2000) - see Section 5. They also note that in the subpolar gyre region of both their model and the observations, the changes in SST lead those in the atmosphere. This is not necessarily an indication that the ocean is driving the atmosphere, since in the model, it is a consequence of the role of ocean dynamics in the evolution of the SST anomalies, and, in particular, of the dynamical adjustment of the ocean model to the specified NAO-forcing.

Eden and Jung's results show interdecadal changes in the horizontal gyre transport of the subtropical and subpolar gyres. Bottom pressure torque, associated with the deep circulation in the model, plays an important role in determining the transport changes. Eden and Willebrand(2000) show an example in which changes in the transport of the model's subpolar gyre are driven by changes in bottom pressure torque associated with the anomalous heat flux forcing used to drive the model, the gyre transport lagging the heat flux forcing by about 2-3 years. There is also evidence that the transport of the Gulf Stream undergoes changes on the interannual and interdecadal time scales (Worthington, 1977; Greatbatch et al., 1991; Sato and Rossby, 1995). The diagnostic calculations of Greatbatch et al.(1991) suggest that the Gulf Stream was reduced in transport by about a third in the early 1970's compared to the late 1950's, a change that is much larger than is found in Eden and Jung's model. Ezer, Mellor and Greatbatch(1995), using the Princeton Ocean Model, diagnosed a similar change in transport to that found by Greatbatch et al.(1991) and were also able to show that the model predicted changes in coastal sea level agree quite well with the observed changes in sea level between the late 1950's and early 1970's. Although changes in the surface wind stress play some role in leading to the model-computed changes in sea level, the dominant influence is that of the change in the large scale ocean circulation computed by the model. In the diagnostic models of Greatbatch et al. and Ezer et al., the change in gyre transport is almost entirely due to the bottom pressure torque, implying a role for the deep circulation.

The NAO can influence the THC through its influence on deep water renewal. Deep convection, and the associated renewal of the ocean's dense water masses, does not occur in every year. For example, Lazier(1980) has pointed out that deep convection did not occur in the Labrador Sea during the late 1960's and early 1970's, a time when the NAO index was predominantly low (Figure 1). In fact, the occurrence of deep convection in the Labrador Sea depends strongly on the severity of the winter in eastern Canada, which in turn is linked to the NAO (Ikeda, 1990; Dickson et al., 1996). When the NAO index is high, the westerly winds are stronger than normal and there are more frequent cold, dry air outbreaks from Labrador. These cold air outbreaks cool the surface layer of the Labrador Sea and, in late winter, can lead to strong convective overturning, sometimes to a depth of over 2000m (Marshall and Schott, 1999). On the other hand, when the NAO is low, the westerly winds are weaker than normal, and outbreaks of cold, dry air from the continental interior are less frequent and less severe. The likelihood of deep water renewal is then correspondingly reduced, as happened in the late 1960's. Dickson et al.(1996) point out that the influence of the NAO works in the opposite fashion in the Greenland Sea. In particular, when the NAO index is high, the Greenland Sea experiences less severe winter conditions than normal, in contrast to the Labrador Sea, leading to reduced deep convective activity. The opposite is true when the NAO index is low. In the early 1990's, the NAO index reached record high levels. Deep convection in the Labrador Sea penetrated to over 2000m depth, and the newly formed Labrador Sea Water (LSW) was the freshest, coldest, and also the densest ever observed (see Figure 10). By contrast, deep water renewal in the

Greenland Sea was substantially reduced at this time. The seesaw between the Labrador and Greenland Seas is another example of the seesaw in winter temperatures between western Greenland and Europe noted by van Loon and Rogers(1978). The NAO also influences the production of 18^oC mode water in the Sargasso Sea. When the index is low, there are more frequent cold dry air outbreaks from the continental US, leading to more 18^oC mode water formation, the reverse being true when the index is high (see Dickson et al.(1996) for more discussion).

It is of considerable interest to investigate the influence of changes in deep water production on the rest of the North Atlantic. Read and Gould(1992) have traced the influence of the shutdown in the production of LSW in the late 1960's on the thermohaline (subsurface) structure of the North Atlantic. Curry et al.(1998) relate changes in the thermohaline structure to the NAO by noting the influence of the NAO on the production of LSW, and, in turn, the influence of LSW on the thermohaline structure (see also Sy et al., 1997). Curry et al. estimate a time scale of 6 years between the formation of the LSW in the Labrador Sea and its appearance at Bermuda. Molinari et al. (1999) have noted the appearance at 26^oN of the anomalously cold and fresh LSW formed during the high NAO index years of the 1980's. Molinari et al. note that the principal conduit between the formation region of LSW and 26^oN is the Deep Western Boundary Undercurrent. They estimate that the time taken for newly formed LSW to reach 26^oN is about 10 years.

There has also been considerable interest in the occurrence of low salinity anomalies that propagate around the subpolar gyre of the North Atlantic. The most famous example is the Great Salinity Anomaly (GSA) of the late 1960's and 1970's (Dickson et al., 1988), but there have been other instances, as discussed by Belkin et al.(1998). Reverdin et al.(1997) carried out an an EOF analysis on lagged time series of upper ocean salt and heat content and found that the first EOF for salt content accounts for 70% of the variance. This EOF corresponds to salinity anomalies that propagate cyclonically around the Labrador Sea and then northeastwards to Europe on a time scale of 5-10 years, behaviour that is very similar to that of the GSA. The principal component time series shows strong decadal variability (although it should be noted that only 39 years of data were available for the analysis). Reverdin et al. demonstrate the close connection between this propagating salinity mode and a pattern of atmospheric variability closely resembling the NAO (see Figure 17 in their paper). Like Belkin et al., Reverdin et al. note that the low salinity, GSA events, likely have their origin in either the Arctic Ocean or the Canadian Arctic. Reverdin et al. point out the important role played by the slope currents around the Labrador Sea (the West Greenland Current and the Labrador Current) in transporting pulses of fresh water from the Arctic to the rest of the North Atlantic. Häkkinen(1993) carried out a model study to investigate the origin of the GSA. She concluded that the anomalous northerly winds along the east coast of Greenland associated with the low NAO index in the 1960's were important for generating the GSA, confirming an hypothesis of Aagaard and Carmack(1989) that the GSA resulted from increased export of sea-ice and freshwater from the Arctic Ocean via Fram Strait. A slightly different hypothesis, involving an Arctic climate cycle, has been put forward by Mysak et al.(1990). The importance of the atmospheric forcing (and hence the NAO) for generating the interannual variability in sea-ice cover over the Newfoundland and Labrador Shelves has been demonstrated by Ikeda et al.(1988). In particular, severe winter conditions in Labrador, in association with the high index state of the NAO, lead to heavy ice years. The impact of the NAO on the ocean climate of the Newfoundland shelf has also been discussed by Myers et al.(1988).

The important role of the NAO for determining temperature variability in the North Sea has been shown by Dippner(1997a). Heyen and Dippner(1998) show the importance of river runoff associated with precipitation anomalies for determining salinity anomalies in the German Bight, confirming an earlier conclusion of Schott(1966). As Heyen and Dippner(1998) point out, although there is a clear link between the salinity anomalies and the atmospheric circulation, the correlation with the NAO is weak (the SLP pattern most closely associated with the salinity variability is similar to the Scandinavian pattern discussed by Rogers, 1990). Dippner(1997a,b) has shown that the recruitment of certain fish stocks in the North Sea is strongly influenced by the NAO through the influence of the NAO on SST. For the case of cod, low SST (low NAO) favours higher recruitment, as happened in the 1960's. Mann and Drinkwater(1995) have argued that the anomalous ocean conditions associated with the high NAO index in the early 1990's played a role in the collapse of the northern cod stock off the east coast of Newfoundland and Labrador. The ecological impact of salinity anomalies (and hence the NAO) on the Northwest Atlantic has also been discussed by Mertz and Myers(1994), Myers et al.(1993) and the references therein.

4. Some thoughts on the dynamics of the NAO

As pointed out by Barnett(1985), the NAO is an internal mode of variability of the atmospheric circulation. Indeed, low frequency (interannual and interdecadal) variability of the NAO is found in simplified atmospheric circulation models that have only seasonal forcing (James and James, 1989), as well as in sophisticated atmospheric general circulation models (AGCM's) with specified climatologically varying SST at the lower boundary. It follows that dynamical coupling with the ocean is not necessary to understand the basic dynamics underlying the NAO. A recent example is provided by Limpasuvan and Hartmann(1999). These authors have analysed output from a 100 year run of an AGCM with seasonal forcing only and demonstrate the very important role played by the transient and stationary eddy fluxes in the maintenance of the different phases of the AO. Limpasuvan and Hartmann(1999) compute the divergence of the Eliassen-Palm flux and show that the eddies systematically drive the anomalous flow in the upper troposphere in the high and low index states in the model (Figure 11). The role of eddy fluxes in maintaining the NAO has also been noted by Hurrell(1995a) and

Hurrell and van Loon(1997). These authors use European Centre for Medium Range Weather Forecasting (ECMWF) global analyses of atmospheric data to show that the eddy flux of vorticity systematically acts to reinforce and maintain the anomalous atmospheric circulation associated with the high NAO index state, while the eddy flux of heat acts to damp the temperature anomalies that are being generated by the anomalous mean flow. Derome, Brunet and Wang(2000) have analysed potential vorticity fluxes on isentropic surfaces in the upper troposphere and find that the NAO is essentially a free mode of the system, maintained by the eddy fluxes. All these studies point to the conclusion that the AO (and hence the NAO) is fundamentally an atmospheric phenomenon, and that to understand the dynamics of the AO/NAO we must understand eddy/mean flow interaction in the atmosphere. It follows that the AO/NAO is ultimately associated with the nonlinear transfer of energy from the synoptic and other intraseasonal time scales to lower frequencies. For this reason, the AO/NAO likely has a large random component in its variability, a point of view supported by the observation of Wunsch(1999) that the spectrum of the NAO is almost white (see Figure 5).

Although dynamic coupling with the ocean is not necessary to understand the basic dynamics underlying the NAO, many authors have suggested that interdecadal climate variability over the North Atlantic (and hence the NAO) may be a dynamically coupled ocean-atmosphere phenomenon (e.g. Deser and Blackmon, 1993; Wohlleben and Weaver, 1995; Sutton and Allen, 1997; McCartney, 1997; Curry et al., 1998; Grötzner et al., 1998; Timmermann et al., 1998; Marshall et al, 2000). This idea is appealing because the typical limit of predictability for the atmosphere is about three weeks. If the large heat capacity of the ocean compared to the atmosphere can be invoked to extend the memory of the North Atlantic ocean/atmosphere system, then it follows that the NAO may be predictable on much longer time scales extending out to seasonal and beyond (e.g. Griffies and Bryan, 1996). There are at least three difficulties with the argument that variability in SST regulates the variability of the NAO. First there is the controversy that surrounds the atmospheric response to North Atlantic SST anomalies. Related to this controversy is the difficulty in interpreting the results from AGCM's forced by the observed time history of SST and sea-ice, as pointed out by Bretherton and Battisti(2000), a problem that is particularly pertinent to the NAO (see the following discussion). Thirdly there is the growing body of evidence that the stratosphere is an important player in determining the variability of the AO (and hence the NAO) during the winter season. We shall discuss each of these in turn, together with the implications for AO/NAO predictability.

The controversy that surrounds the atmospheric response to midlatitude SST anomalies, and, in particular North Atlantic SST anomalies, is nicely reviewed by Kushnir and Held(1996) and Lau(1997). Some AGCM's show a relatively strong response to North Atlantic SST anomalies (e.g. Rodwell et al., 1999; Mehta et al., 2000), while others show no response at all (e.g. Zwiers et al., 2000), and others show a response that depends on the basic state of the atmospheric circulation (Peng et al., 1995). (It should be noted that the statistical robustness of Peng et al.'s result is not clear). In the studies of Rodwell et al.(1999) and Mehta et al.(2000) an ensemble of AGCM experiments was carried out, each experiment being forced by the time series of observed SST and sea-ice, but each started with a different initial condition. These authors find that the NAO for the ensemble mean of the different runs is highly correlated with the observed NAO index, the correlation being as high as 0.7 on time scales longer than about 6 years. Bretherton and Battisti(2000; hereafter BB) have questioned the interpretation that should be put on the results of Rodwell et al. and Mehta et al.. They point out that the AO/NAO is an important local forcing for North Atlantic SST. This means that even if the overlying atmosphere is sensitive to the underlying SST, there is still the problem of predicting the SST, a problem that is made doubly difficult when, as for the NAO, the overlying atmosphere contains a large, random component that is driving the SST anomalies.

It is worth exploring Bretherton and Battisti's work in more detail. The model is very simple and one-dimensional (the details can be found in Barsugli and Battisti, 1998). The model is formulated in terms of temperature anomalies T_a and T_o for the middle troposphere and the ocean mixed layer, respectively, and is driven by a white noise forcing N representing the chaotic dynamics generated by the synoptic variability. The governing equations are

$$\frac{dT_a}{dt} = -aT_a + bT_o + N \tag{1}$$

and

$$\beta \frac{dT_o}{dt} = cT_a - dT_o. \tag{2}$$

β represents the large heat capacity of the ocean mixed layer compared to the atmosphere and in BB has value 40. The other parameters are a=1.12, b=0.5, c=1, d=1.08 (note that since $b \neq 0$, some sensitivity of the overlying atmosphere to the SST is allowed). The same analysis procedure is then applied to this simple coupled model as was used by Rodwell et al. and Mehta et al. to analyse their experiments. First a time series of the ocean temperature, T_o^C, is generated using a particular realisation of the random number generator N. (We call this the truth run). The time series T_o^C is then used to drive equation (1), i.e. the atmospheric model, in an ensemble of experiments with different realisations of N. The results in Table 1 show that this very simple model is able to reproduce the behaviour found by Rodwell et al. and Mehta et al., including the high correlation between the atmospheric temperature of the ensemble mean and the atmospheric temperature in the truth run. This is true even though T_o^C is generated in the truth run by random forcing from the atmosphere model. The reason for this behaviour is that the ensemble averaging acts to filter out the noise from the atmosphere only runs, leaving only the signal associated with the SST (that part of the solution to (1) that arises because $b \neq 0$). However, because the particular time series T_o^C is itself generated by the random forcing from the atmosphere model, there is no predictability within the coupled system beyond that associated with the large heat capacity of the ocean. Indeed, BB estimate that despite the excellent hindcast skill of an atmospheric ensemble forced by observed SST's, less than 15% of the variance in the seasonal atmospheric (NAO) anomaly is predictable six months in advance, even given perfect initial conditions for the midlatitude ocean state. The simple model also explains what was a puzzling feature of Rodwell et al.'s results, namely that the computed air/sea fluxes act to damp the SST anomalies in Rodwell et al.'s experiments, not reinforce them, as happens in nature (see Section 3). This is also a feature of the simple linear model and arises because in the atmosphere-only experiments, the specified SST is a driving force for the atmosphere model and tends to lead, rather than lag, the atmospheric temperature.

The difficulty pointed out by BB arises when the SST in the coupled system is being driven by atmospheric variability that has a large random component. The models of Rodwell et al. and Mehta et al. nevertheless do have some response to the underlying SST. In BB's model this is taken into account by the nonzero value for b. It follows that if SST is influenced by some process other than local random forcing from the atmosphere, dynamic coupling between the atmosphere and the ocean might lead to some significant predictability. We have seen that on interdecadal time scales (30 years and longer), North Atlantic SST can evolve independently of the local forcing (Bjerknes, 1964), as recently demonstrated by Eden and Jung(2000). It is therefore possible that some information about the future (interdecadal time scale) atmospheric state may reside in the slow, interdecadal evolution of the ocean. The possibility of predictive skill within the ocean on decadal time scales has been noted by Griffies and Bryan(1996). It should be cautioned, however, that the percentage of the atmospheric variance that might be predictable in this way may turn out to be very small, as implied by the almost white spectrum of the NAO noted when discussing Figure 5 (Wunsch, 1999). One other possibility still under investigation is that SST in the tropical Atlantic may play a role in modulating the variability of the AO/NAO, perhaps through the TAV. (TAV is Tropical Atlantic Variability: see Nobre and Shukla(1996) and also the Atlantic Climate Variability Experiment (ACVE) prospectus at http://www.ldeo.columbia.edu/ visbeck/acve/report/acve_report.html.) It is known that the atmosphere is much more sensitive to the underlying SST in the tropics than at midlatitudes (Lau, 1997). This possibility is suggested by the work of Robertson et al.(2000), who also find an influence of South Atlantic SST on the NAO in their model. There is also the possibility that the atmospheric circulation may exhibit a more significant response to the underlying SST at one time of year than at another, as suggested by Peng et al.(1995). Czaja and Frankignoul(1999) find evidence of a significant atmospheric response in the spring to SST anomalies in the previous winter, and also in the early winter in response to SST anomalies the previous summer. Since the SST anomalies persist for several months, it is possible that while the atmospheric flow may be insensitive to the underlying SST when the anomalies are generated, there may be greater sensitivity as the atmospheric flow evolves to a different state under the seasonal forcing.

As noted in Section 2, the AO is an equivalent barotropic feature of the atmospheric circulation whose energy density in winter is almost invariant with height between the 1000-hPa and 50-hPa levels (Thompson and Wallace, 1998). Indeed, fluctuations in the strength of the wintertime stratospheric vortex are strongly linked to SLP variability in both the northern and southern hemispheres (Thompson and Wallace, 2000). Baldwin et al.(1994), Perlwitz and Graf(1995), Cheng and Dunkerton(1995), Kitoh et al.(1996) and Kodera et al.(1996) have all documented coupling between the the winter stratospheric polar vortex and its tropospheric counterpart. This suggests a possible role for the stratosphere in modulating the AO/NAO. Recently, Baldwin and Dunkerton(1999) have shown that AO anomalies during wintertime often first appear in the stratosphere near the 10-hPa level and then propagate downward into the troposphere on a time scale of weeks (see Figure 12). They show that the midwinter correlation between the 90 day low pass filtered 10 hPa and 1000 hPa AO-anomalies exceeds 0.65 when the surface anomaly time series is lagged by about three weeks. It should be possible, therefore, to use information from the stratosphere to improve prediction of the AO/NAO in the troposphere one month ahead during the winter season, with the implication that models used for long-range prediction may require a detailed representation of the stratosphere. In Section 2, it was noted that the stratosphere may play a role in explaining the increasing trend of the AO/AAO that is found in some recent model experiments under greenhouse gas forcing (Shindell et al., 1999; Fyfe et al., 1999; Paeth et al., 1999). Clearly, more research is required to unravel the dynamic link between the stratosphere and AO/NAO, a topic that is of interest on a wide range of time scales.

The importance of wave mean flow interaction for the dynamics of the AO/NAO suggests a link between the NAO and the frequency of atmospheric blocking. A blocking pattern, such as shown in Figure 13, projects negatively on the NAO index (the pressure difference between Iceland and the Azores) and the recurrence of such events throughout a season would lead to a low NAO index. Hurrell(1996) notes that the 1960's, when the NAO index was low, was a period of frequent wintertime blocking over the North Atlantic. Huang et al.(2000) have shown that both the frequency and lifetime of blocking events in the North Atlantic region is related to the phase of the NAO, with more frequent, longer lasting blocks when the index is low. Vautard(1990) has also noted a link between blocking over Greenland (his Greenland Anticyclone weather regime) and the NAO. What is not clear is whether the blocks are playing a role in causing periods of low index or whether they are merely a symptom of the weakened polar vortex associated with low index. Blocking events in the troposphere are known to be linked with stratospheric sudden warmings (Andrews et al., 1987; Quiroz, 1986) suggesting, once again, a dynamical link between the NAO and stratosphere.

5. The "Null Hypothesis" for North Atlantic interannual/interdecadal variability

The discussion in Section 4 suggests that dynamical coupling between the atmosphere and ocean is not strong over the North Atlantic, at least on time scales out to interannual and interdecadal. It follows that the simplest possible interpretation for interannual to interdecadal variability in the North Atlantic climate system is the "Null Hypothesis", according to which the midlatitude ocean responds passively to the overlying atmospheric variability, and, in particular, does not feed back and dynamically influence the overlying atmospheric circulation. (For more discussion of the "Null Hypothesis" see the Atlantic Climate Variability Experiment prospectus at http://www.ldeo.columbia.edu/ visbeck/acve/report/acve_report.html.) This view is consistent with Dickson et al.(1996) who state "we see strong evidence of a direct impact of the shifting atmospheric circulation on the ocean; while this certainly does not rule out either feedbacks from anomalous ice and SST conditions on the atmosphere, or autonomous oscillations of the ocean's overturning circulation, it does tend to minimize them".

Delworth and Greatbatch(2000) describe an application of the "Null Hypothesis" to understand the interdecadal variability in the Atlantic sector of the GFDL coupled ocean/atmosphere model (Delworth et al., 1993). Figure 14 shows the times series of the strength of the thermohaline circulation (THC) in the coupled model. To understand the variability, we begin by noting that the ocean component of the coupled model does not support its own self-sustained interdecadal variability in the Atlantic sector (Greatbatch et al., 1997). This is because the realistic bottom topography acts as a strong damping (Winton, 1997), and contrasts with the self-sustained interdecadal variability found in the flat-bottomed models of Weaver and Sarachik(1991), Greatbatch and Zhang(1995) and many similar studies. Rather, the THC variability in the coupled model is sustained by the variable atmospheric forcing, as in the box model of Griffies and Tziperman(1995). Delworth and Greatbatch(2000) show that the surface heat flux is the dominant term (relative to the fresh water and momentum fluxes) in driving the THC variability (similar to the finding of Eden and Jung, 2000) and also that the interdecadal THC variability in the coupled model is driven by the low frequency portion of the spectrum of atmospheric flux forcing (time scales of 20 years and longer). Analyses also reveal that the THC fluctuations are driven by a spatial pattern of surface heat flux variations that bears a strong resemblance to the North Atlantic Oscillation (Figure 15).

To assess the importance of dynamic coupling between the atmosphere and ocean models, the results of two experiments are described in which the coupling is suppressed (Figure 16). In RANDOM, the ocean component of the coupled model is driven by annual mean surface fluxes taken from the coupled model but randomly rearranged in time, and in ATMOS, the ocean model is driven by a time series of annual mean fluxes taken from the atmospheric component of the coupled model run with specified seasonally varying SST and sea-ice. Both model runs show that significant interdecadal variability in the THC can be generated by low frequency atmospheric forcing that does not rely on dynamical coupling with the ocean. In ATMOS, the interdecadal variability in the THC is being driven entirely by interdecadal variability generated internally within the atmospheric model. In RANDOM, the flux forcing has a white spectrum and is varying completely independently of the underlying SST. The amplitude of the THC variability in RANDOM and ATMOS is nevertheless reduced compared to that in the coupled model, particularly in ATMOS (the standard deviation of the THC variability in the coupled model is 0.98 Sv, compared with 0.77 Sv in RANDOM and 0.47 Sv in ATMOS). Delworth and Greatbatch(2000) argue that this is because thermodynamic coupling (as well as dynamic coupling) is suppressed in ATMOS and RANDOM. Thermodynamic coupling is important in the deep convection region where heat can be stored in the ocean during periods of reduced convection and then released when convection becomes active again, a process that leads to enhanced heat flux variability in the deep convection region. The absense of any influence from deep convection is very evident in ATMOS and can be seen by comparing the three panels in Figure 15. (In the model, deep convection takes place primarily to the south of Greenland.) Delworth and Greatbatch(2000) conclude that there is no evidence that the THC variability in the coupled model is part of a dynamically coupled mode of the atmosphere and ocean models. We believe, therefore, that the THC variability in the GFDL model is consistent with the “Null Hypothesis”.

6. Summary and Conclusions

The NAO is the most important mode of variability in the atmospheric circulation over the North Atlantic, with considerable influence on winter temperature throughout the Eurasian continent and eastern North America (Figure 3). The NAO is closely related to the AO which, as pointed out by Thompson and Wallace(1998), is a hemispheric mode of variability that extends throughout the depth of the troposphere and up into the winter stratosphere. Indeed, there is growing evidence that the stratosphere plays an important role in winter AO/NAO variability (Baldwin and Dunkerton, 1999) and may also be implicated in the upward trend of the AO index that is a feature of some model simulations that include greenhouse gas forcing (Shindell et al., 1999; Fyfe et al., 2000; Paeth et al., 1999). The AO/NAO is also an important forcing for the North Atlantic Ocean (Section 3). AO/NAO driving of North Atlantic SST anomalies is implicit in the classic study of Bjerknes(1964), and reiterated by Cayan(1992) and many subsequent studies. Curry et al.(1998) point out that the signature of the NAO is felt throughout the surface and subsurface waters of the North Atlantic thermocline, and recent modelling studies (Häkkinen, 1999; Eden and Jung, 2000; Eden and Willebrand, 2000; Delworth and Greatbatch, 2000) emphasise the importance of the AO/NAO for driving circulation changes in the North Atlantic, in particular the thermohaline overturning circulation (THC).

Section 4 provided some discussion on the dynamics of the NAO. It was emphasised that the AO/NAO is an internal mode of variability of the atmospheric circulation driven by eddy/mean flow interaction (e.g. Limpasuvan and Hartmann, 1999), and that dynamical coupling with the ocean is probably weak. Recent studies by Rodwell et al.(1999) and Mehta et al.(2000) have used an ensemble of atmospheric general circulation model (AGCM) experiments, driven by the observed time series of SST and sea-ice, to reproduce the observed evolution of the NAO, apparently holding out the prospect of predicting the NAO several years in advance. Bretherton and Battisti(2000) have questioned this interpretation. They point out that to predict the NAO, one must first predict the SST and sea-ice, a problem that is very difficult when, as for the NAO, the atmosphere contains a large random component that is itself an important local forcing for SST and sea-ice. The difficulty pointed out by Bretherton and Battisti(2000) can be circumvented when some process other than local atmospheric forcing is driving the SST, an example being changes in oceanic heat transport (see Figure 9 and the discussion thereon). On the other hand, the spectrum of the winter NAO index (Figure 5) is only weakly red, with no striking peaks (Wunsch, 1999), suggesting that even on the interdecadal time scale, there is little hope for prediction of the NAO, a conclusion that, nevertheless, would benefit from further study using coupled ocean/atmosphere models. More exciting from the predictability point of view is the possibility, suggested by the study of Baldwin and Dunkerton(1999) (see Figure 12), that information in the winter stratosphere can be used to provide information on the likely phase of the AO/NAO in the lower troposphere one month ahead. The relationship between the NAO and quasi-stationary regimes of the atmospheric circulation (Vautard, 1990; Monahan et al., 2000) also requires further investigation. Sometimes the atmospheric circulation remains close to one of these regimes for extended periods, exceeding the typical three week limit of predictability. It follows that understanding these regimes and transitions between them could prove helpful for improving prediction one season ahead.

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References

Aagaard K,Carmack EC(1989) The role of sea-ice and other freshwater in the Arctic circulation. J Geophys Res 94:14485-14498

Andrews DG,Holton JR,Leovy CB(1987) Middle Atmosphere Dynamics. Academic Press, 489 pp.

Appenzeller C,Stocker TF,Anklin M(1998) North Atlantic Oscillation dynamics recorded in Greenland ice cores. Science 282:446-449

Baldwin MP,Cheng X,Dunkerton TJ(1994) Observed correlations between winter-mean tropospheric and stratospheric circulation anomalies. Geophys Res Lett 21:1141-1144

Baldwin MP,Dunkerton TJ(1999) Propagation of the Arctic Oscillation from the stratosphere to the troposphere. J Geophys Res 104:30937-30946

Barnett TP(1985) Variations in near-global sea level pressure. J Atmos Sci 42:478-501

Barnston AG, Livezey RE(1987) Classification, seasonality and persistence of low frequency atmospheric circulation patterns. Mon Wea Rev 115:1083-1126

Barsugli JJ,Battisti DS(1998) The basic effects of atmosphere-ocean thermal coupling on midlatitude variability. J Atmos Sci 55:477-493

Battisti DS,Bhatt US,Alexander MA(1995) A modelling study of the interannual variability in the wintertime North Atlantic Ocean. J Climate 8:3067-3083

Belkin IM,Levitus S,Antonov J,Malmberg S-A(1998) “Great Salinity Anomalies” in the North Atlantic. Prog Oceanogr 41:1-68

Bjerknes J(1964) Atlantic air-sea interaction. Adv Geophys 10:1-82

Bretherton CS,Battisti DS(2000) An interpretation of the results from atmospheric general circulation models forced by the time history of the observed sea surface temperature distribution. Geophys Res Lett 27:767-770

Broecker WS(1991) The great ocean conveyor. Oceanography 4:79-89

Cayan DR(1992) Latent and sensible heat flux anomalies over the northern oceans: Driving the sea surface temperature. J Phys Oceanogr 22:859-881

Cheng X,Dunkerton TJ(1995) Orthogonal rotation of spatial patterns derived from singular value decomposition analysis. J Climate 8:2631-2643

Cook ER,D'Arrigo RD,Briffa KR(1998) The North Atlantic Oscillation and its expression in circum-Atlantic tree-ring chronologies from North America and Europe. The Holecene 8:9-17

Cullen H, D'Arrigo R,Cook E,Mann ME(2000) Multiproxy reconstructions of the North Atlantic Oscillation. Paleoceanography submitted

Curry RG,McCartney MS,Joyce TM(1998) Oceanic transport of subpolar climate signals to mid-depth subtropical waters. Nature 391:575-577

Czaja A,Frankignoul C(1999) Influence of the North Atlantic SST on the atmospheric circulation. Geophys Res Lett 26:2969-2972

Delworth TL(1996) North Atlantic interannual variability in a coupled ocean-atmosphere model. J Climate 9:2356-2375

Delworth TL,Greatbatch RJ(2000) Multidecadal thermohaline circulation variability driven by atmospheric surface flux forcing. J Climate 13:1481-1495

Delworth TL,Manabe S,Stouffer RJ(1993) Interdecadal variations of the thermohaline circulation in a coupled ocean-atmosphere model. J Climate 6:1993-2011.

Derome J,G Brunet,Y Wang(2000) On the potential vorticity balance on an isentropic surface during normal and anomalous winters. Mon Wea Rev submitted

Deser C(2000) A note on the teleconnectivity of the "Arctic Oscillation". Geophys Res Lett 27:779-782

Deser C,Blackmon ML(1993) Surface climate variations over the North Atlantic Ocean during winter: 1900-1989. J Climate 6:1743-1753

Dickson RR,Lazier J,Meincke J,Rhines P,Swift J(1996) Long-term coordinated changes in the convective activity of the North Atlantic. Prog Oceanogr 38:214-295

Dickson RR,Meinke J,Malmberg S-A, Lee AJ(1988) The "Great Salinity Anomaly" in the Northern North Atlantic, 1968-1982. Prog Oceanogr 20:103-151

Dippner, JW(1997a) SST anomalies in the North Sea in relation to the North Atlantic Oscillation and the influence on the theoretical spawning time of fish. Dtsch Hydrogr Z 49:267-275

Dippner, JW(1997b) Recruitment success of different fish stocks in the North Sea in relation to climate variability. Dtsch Hydrogr Z 49:277-293

Eden C,Jung T(2000) North Atlantic interdecadal variability: Oceanic response to the North Atlantic Oscillation (1865-1997). J Climate submitted

Eden C,Willebrand J(2000) Mechanism of interannual to decadal variability of the North Atlantic circulation. J Climate submitted

Ezer T,Mellor GL,Greatbatch RJ(1995) On the interpentadal variability of the North Atlantic Ocean: Model simulated changes in transport, meridional heat flux and coastal sea level between 1955-59 and 1970-1974. J Geophy Res 100:10,559-10,566

Fyfe JC,Boer GJ,Flato GM(1999) The Arctic and Antarctic Oscillations and their projected changes under global warming. Geophys Res Lett 26:1606-1604

Greatbatch RJ,Fanning AF,Goulding A,Levitus S(1991) A diagnosis of interpentadal circulation changes in the North Atlantic. J Geophys Res 96:22009-22024

Greatbatch RJ,Peterson KA,Roth H(1997) Interdecadal variability in a coarse resolution model with North Atlantic bottom topography. Department of Oceanography, Dalhousie University, Halifax, Nova Scotia, Canada.

Greatbatch RJ,Zhang S(1995) An interdecadal oscillation in an idealised ocean basin forced by constant heat flux. J Climate 8:81-91.

Griffies S,Bryan K(1996) Predictability of North Atlantic multidecadal climate variability. Science 275:181-184

Griffies S,Tziperman E(1995) A linear thermohaline oscillator driven by stochastic atmospheric forcing. J Climate 8:2440-2453

Grötzner A,Latif M,Barnett TP(1998) A decadal climate cycle in the North Atlantic Ocean as simulated by the ECHO coupled model. J Climate 11:831-847

Häkkinen S(1993) An Arctic source for the Great Salinity Anomaly:A simulation of the Arctic ice-ocean system for 1955-1975. J Geophys Res 98:16397-16410

Häkkinen S(1999) Variability of the simulated meridional heat transport in the North Atlantic for the period 1951-1993. J Geophys Res 104:10,991-11,007

Hall MM,Bryden H(1982) Direct estimates and mechanisms of ocean heat transport. Deep

Sea Res Part A 29:339-359

Halliwell GR(1997) Decadal and multidecadal North Atlantic SST anomalies driven by standing and propagating basin-scale atmospheric anomalies. J Climate 10:2405-2411

Halliwell GR(1998) Simulation of North Atlantic decadal/multidecadal winter SST anomalies driven by basin-scale atmospheric circulation anomalies. J Phys Oceanogr 28:5-21

Halliwell GR,Mayer DA(1996) Frequency response properties of forced climate SST anomaly variability in the North Atlantic. J Climate 9:3575-3587

Hansen DV,Bezdek H(1996) On the nature of decadal anomalies in North Atlantic sea surface temperature. J Geophys Res 101:8749-8758

Heyen H,Dippner JW(1998) Salinity variability in the German Bight in relation to climate variability. Tellus 50A:545-556

Huang J,Shabbar A,Higuchi K(2000) The relationship between the wintertime North Atlantic Oscillation and blocking episodes in the North Atlantic. Int J Climatol submitted

Hurrell JW(1995a) Transient eddy forcing of the rotational flow during northern winter. J Atmos Sci 52:2286-2301

Hurrell JW(1995b) Decadal trends in the North Atlantic Oscillation: Regional temperatures and precipitation. Science 269:676-679

Hurrell JW(1996) Influence of variations in extratropical wintertime teleconnections on Northern Hemisphere temperature. Geophys Res Lett 23:665-668

Hurrell JW,van Loon H(1997) Decadal variations in climate associated with the North Atlantic Oscillation. Climate Change 36:301-326

Ikeda M(1990) Decadal oscillations of the air-ice-ocean system in the Northern hemisphere. Atmos.-Ocean 28:106-139

Ikeda M,Symonds G,Yao T(1988) Simulated fluctuations in annual Labrador Sea ice cover. Atmos-Ocean 26:16-39

James IN, James PM(1989) Ultra-low-frequency variability in a simple atmospheric circulation model. Nature 342:53-55

Kalnay E, et al.(1996) The NCEP/NCAR 40 year reanalysis project. Bull Amer Met Soc 77:437-471

Kitoh A,Koide H,Kodera K,Yukimoto S,Noda A(1996) Interannual variability in the stratospheric-tropospheric circulation in a coupled ocean-atmosphere GCM. Geophys. Res Lett 23:535-546

Kodera K,Chiba M,Koide H,Kitoh A,Nikaidou Y(1996) Interannual variability of the winter stratosphere and troposphere in the Northern hemisphere. J Met Soc Japan 74:365-382

Kushnir Y(1994) Interdecadal variations in North Atlantic sea surface temperature and associated atmospheric conditions. J Climate 7:141-157

Kushnir Y(1999) Europe's winter prospects. Nature 398:289-291

Kushnir Y,Held IM(1995) Equilibrium atmospheric response to North Atlantic SST anomalies. J Climate 9:1208-1220.

Lau N-C(1997) Interactions between global SST anomalies and the midlatitude atmospheric circulation. Bull Amer Met Soc 78:21-33

Lazier JRN(1980) Oceanic conditions at Ocean Weather Ship Bravo, 1964-1974. Atmos-Ocean, 18:227-238

Limpasuvan V,Hartmann DL(1999) Eddies and the annular modes of climate variability. Geophys Res Lett 26:3133-3136

Loewe F(1937) A period of warm winters in western Greenland and the temperature seesaw between western Greenland and central Europe. Quart J Roy Met Soc 63:365-372

Loewe F(1966) The temperature seesaw between western Greenland and Europe. Weather 21:241-246

Luterbacher J,Schmutz C,Gyalistras D, Xoplaki E(1999) Reconstruction of monthly NAO and EU indices. Geophys Res Lett 26:2745-2748

Mann KH,Drinkwater K(1994) Environmental influences on fish and shellfish production in the Northwest Atlantic. Environ Rev 2:16-32

Marshall J,Johnson H,Goodman J(2000) A study of the interaction of the North Atlantic Oscillation with the ocean circulation. J Climate submitted

Marshall J, Schott F(1999) Open ocean deep convection: observations, models and theory. Rev of Geophysics 37:1-64

McCartney M(1997) Is the ocean at the helm? Nature 388:521-522

Mehta VM,Suarez MJ,Manganello J,Delworth TL(2000) Oceanic influence on the North Atlantic Oscillation and associated northern hemisphere climate variations:1959-1993. Geophys Res Lett 27:121-124

Meinhardus W(1898) Uber einige meteorologische Beziehungen zwischen dem Nordatlantischen Ozean und Europa im Winterhalbjahr. Meteor Z 15:85-105

Mertz G,Myers RA(1994) The ecological impact of the Great Salinity Anomaly in the northern North-west Atlantic. Fish Oceanogr 3:1-14

Molinari RL,Fine RA,Wilson WD,Curry RG,Abell J,McCartney M(1998) The arrival of recently formed Labrador Sea Water in the Deep Western Boundary Current at 26.5^{o}N. Geophys Res Lett 25:2249-2252

Monahan AH,Fyfe JC,Flato GM(2000) A regime view of northern hemisphere atmospheric variability and change under global warming. Geophys Res Lett, submitted.

Mysak LA,Manak DK,Marsden RF(1990) Sea-ice anomalies observed in the Greenland and Labrador Seas during 1901-1984 and their relation to an interdecadal Arctic climate cycle. Climate Dynamics 5: 111-133

Myers RA,Akenhead S,Drinkwater KF(1988) The North Atlantic Oscillation and the ocean climate of the Newfoundland shelf. NAFO SCR Doc 88/65, 22pp

Myers RA,Drinkwater KF,Barrowman NJ,Baird JW(1993) Salinity and recruitment of Atlantic cod(Gadus morhua) in the Newfoundland region. Can J Fish Aquat Sci 50:1599-1609

Nobre P,Shukla J(1996) Variations of sea surface temperature, wind stress, and rainfall over the tropical Atlantic and South America. J Climate 9:2464-2479

Paeth H,Hense A,Glowienka-Hense R,Voss R,Cubasch U(1999) The North Atlantic Oscillation as an indicator for greenhouse-gas induced regional climate change. Clim Dyn 15:953-960

Peng S,Mysak LA,Ritchie H,Derome J,Dugas B(1995) The difference between early and midwinter atmospheric responses to sea surface temperature anomalies in the Northwest Atlantic. J Climate 8:137-157

Perlwitz J,Graf H-F(1995) The statistical connection between tropospheric and stratospheric circulation of the Northern Hemisphere in winter. J Climate 8:2281-2295

Quiroz RS(1986) The association of stratospheric warmings and tropospheric blocking. J Geophys Res 91:5277-5285

Read JF,Gould WJ(1992) Cooling and freshening of the subpolar North Atlantic Ocean since the 1960's. Nature 360: 55-57

Reverdin G,Cayan D,Kushnir Y(1997) Decadal variability of hydrography in the upper northern North Atlantic in 1948-1990. J Geophys Res 102:8505-8532

Robertson AW,Mechoso CR,Kim Y-J(2000) The influence of Atlantic sea surface temperature anomalies on the North Atlantic Oscillation. J Climate 13:122-138

Rodwell MJ,Rowell DP,Folland CK(1999)Oceanic forcing of the wintertime North Atlantic Oscillation and European climate. Nature 398:320-323

Rogers JC(1984) The association between the North Atlantic Oscillation and the Southern Oscillation in the Northern Hemisphere. Mon Wea Rev 112:1999-2015

Rogers JC(1990) Patterns of low frequency monthly sea level pressure variability (1899-1996) and associated wave cyclone frequencies. J Climate 3:1364-1379

Rogers JC,van Loon H(1979) The seesaw in winter temperatures between Greenland and northern Europe. Part II:Some oceanic and atmospheric effects at middle and high latitudes. Mon Wea Rev 107:509-519

Sato O,Rossby T(1995) Seasonal and secular variations in dynamic height anomaly and transport of the Gulf Stream. Deep Sea Res 42:149-164

Shindell DT,Miller RL,Schmidt GA,Pandolfo L(1999) Simulation of recent northern climate trends by greenhouse-gas forcing. Nature 399:452-455

Schott F(1966) Der Oberflächensalzgehalt in der Nordsee. Dtsch Hydrogr Z (suppl A9):1-58

Shutts GJ(1986) A case study of eddy forcing during an Atlantic blocking episode. Adv Geophys 29:135-162

Stockton C, Glueck M(1999) Long-term variability of the North Atlantic Oscillation. In:American Meteorological Society 10th Symposium on Global Change Studies, Dallas Texas, pp.290-293

Sutton RT, Allen MR(1997) Decadal predictability of North Atlantic sea surface temperature and climate. Nature 388:563-567

Sy A,Rhein M,Lazier JRN,Koltermann KP,Meincke J,Putzka A,Bersch M(1997) Surprisingly rapid spreading of newly formed intermediate waters across the North Atlantic Ocean. Nature 386:675-679

Thompson DWJ,Wallace JM(1998) The Arctic Oscillation signature in wintertime geopotential height and temperature fields. Geophys Res Lett 25:1297-1300

Thompson DWJ,Wallace JM(2000) Annular modes in the extratropical circulation. Part I:Month to month variability. J Climate 13:1000-1016

Thompson DWJ,Wallace JM,Hegerl GC(2000) Annular modes in the extratropical ciruclation. Part II:Trends. J Climate 13:1018-1036

Timmermann A,Latif M,Voss R,Grötzner A(1998) Northern hemispheric interdecadal variability: A coupled air-sea mode. J Climate 11:1906-1931.

van Loon H,Rogers JC(1978) The seesaw in winter temperatures between Greenland and northern Europe. Part I:General description. Mon Wea Rev 106:296-310

Vautard R(1990) Multiple weather regimes over the North Atlantic: Analysis of precursors and successors. Mon Wea Rev 118:2056-2081

Visbeck M,Cullen H, Krahmann G,Naik N(1998) An ocean model's response to North Atlantic Oscillation-like wind forcing. Geophys Res Lett 25:4521-4524

Wallace JM,Gutzler DS(1981) Teleconnections in the geopotential height field during the northern hemisphere winter. Mon Wea Rev 109:784-812

Walker GT(1924) Correlations in seasonal variations of weather IX. Mem Ind Meteor Dept 24:275-332

Walker GT,Bliss EW(1932) World weather V. Mem Roy Meteor Soc 4:53-84

Weaver AJ,Sarachik ES(1991) Evidence for Decadal Variability in an ocean general circulation model: An advective mechanism. Atmosphere-Ocean 29:197-231.

Winton M(1997) The damping effect of bottom topography on internal decadal-scale oscillations of the thermohaline circulation. J Phys Oceanogr 26:289-304

Wohlleben TMH, Weaver AJ(1995) Interdecadal climate variability in the subpolar North Atlantic. Climate Dynamics 11:459-467

Worthington LV(1977) The intensification of the Gulf Stream after the winter of 1976-77.

Nature 270:415-417

Wunsch C(1999) The interpretation of short climate records, with comments on the North Atlantic and Southern Oscillations. Bull Amer Met Soc 80:245-255

Zwiers FW,Wang XL,Sheng J(2000) Effects of specifying bottom boundary conditions in an ensemble of atmospheric GCM simulations. J Geophys Res 105:7295-7315

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Table 1. Simulated NAO variability given specified SST in two AGCM studies and the Bretherton and Battisti(BB) model.

Experiment	N^{a}	Seasonal[b] R^{d}	Low Freq.[c] R^{d}	AR[e]
Rodwell et al.	6	0.21 / 0.41	0.43 / 0.74	0.39
Mehta et al.	16	0.17 / 0.43	0.28 / 0.75	0.50
BB Model	16	0.18 / 0.45	0.45 / 0.80	0.41

[a]The number of AGCM realizations in the ensemble

[b]Seasonal variability is loosely defined as the 3-month winter average for *Rodwell et al.*(1999), the one month average for *Mehta et al.*(2000), and 3 months for the BB model.

[c]'Low Freq.' refers to time series low-pass filtered to retain periods greater than 6.5 years.

[d]Correlation coefficients are between the simulated and observed NAO index (AGCMs) or atmospheric temperature anomaly (BB model). The first value is the average for one atmospheric realization, and the second value is for the ensemble mean prediction.

[e]Ratio between standard deviation of low-pass atmospheric variability in the model to that observed.

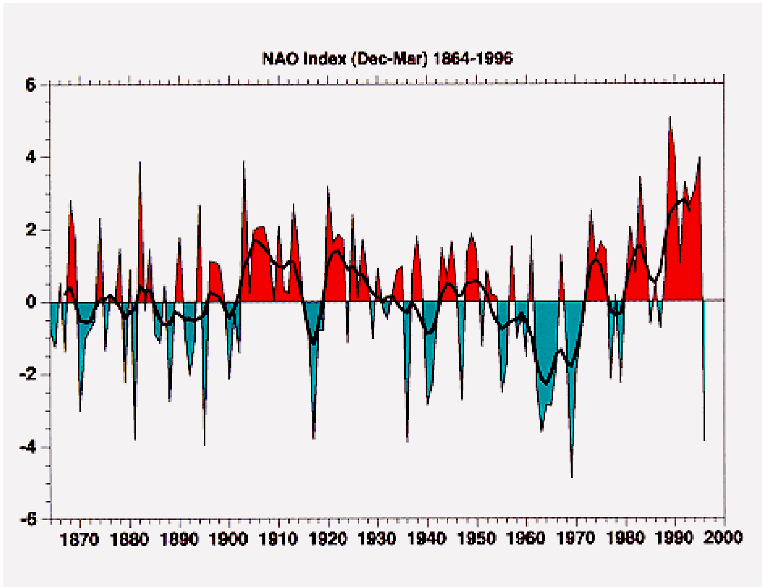

Figure 1. The NAO index from 1864 to 1996 defined as in Hurrell(1995b). The heavy line shows the time series after filtering to remove periods less than 4 years.

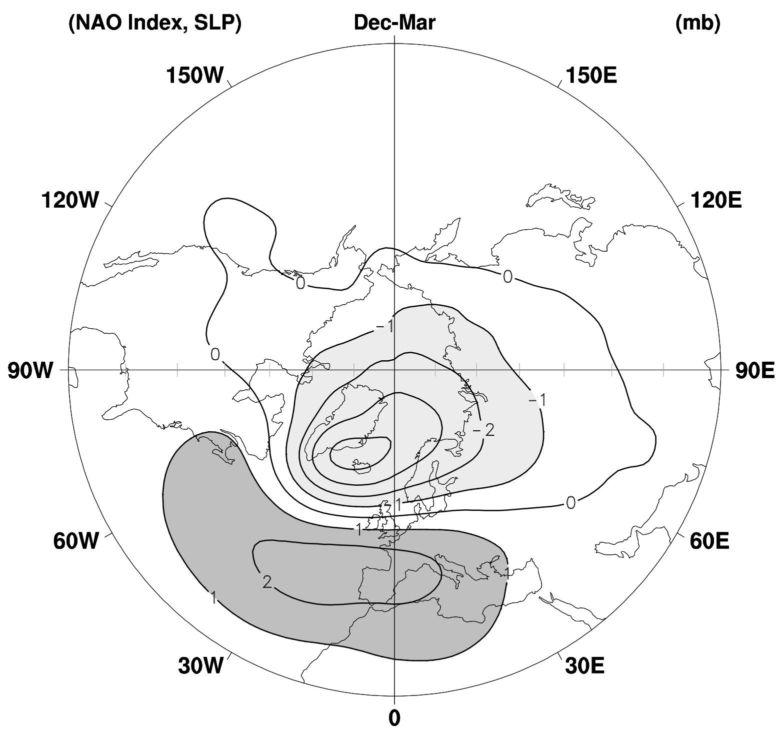

Figure 2. SLP change associated with one standard deviation (positive) of the NAO index. Courtesy of J. Hurrell.

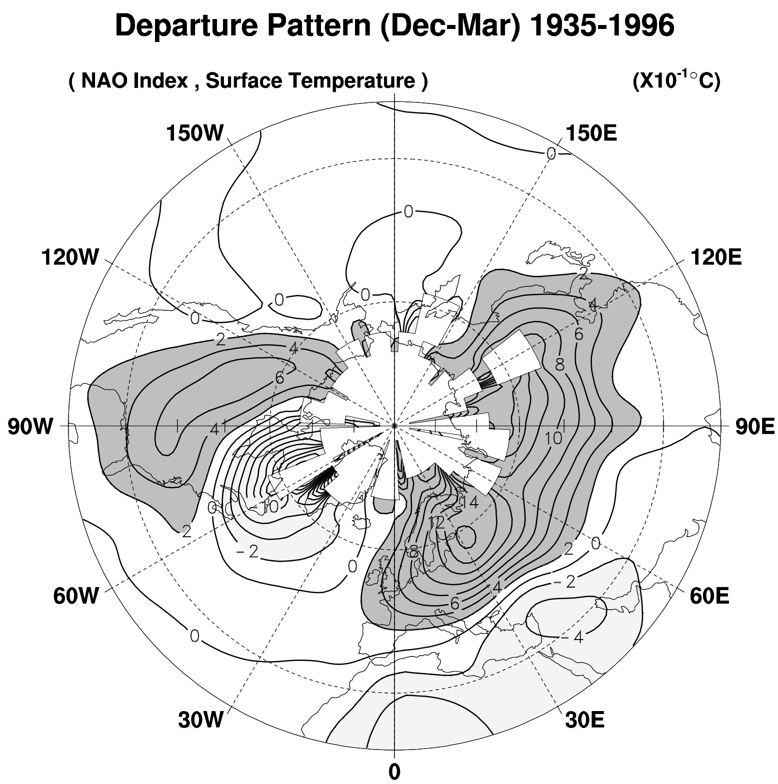

Figure 3. Surface temperature change associated with one standard deviation (positive) of the NAO index. From Hurrell(1996), courtesy of J. Hurrell.

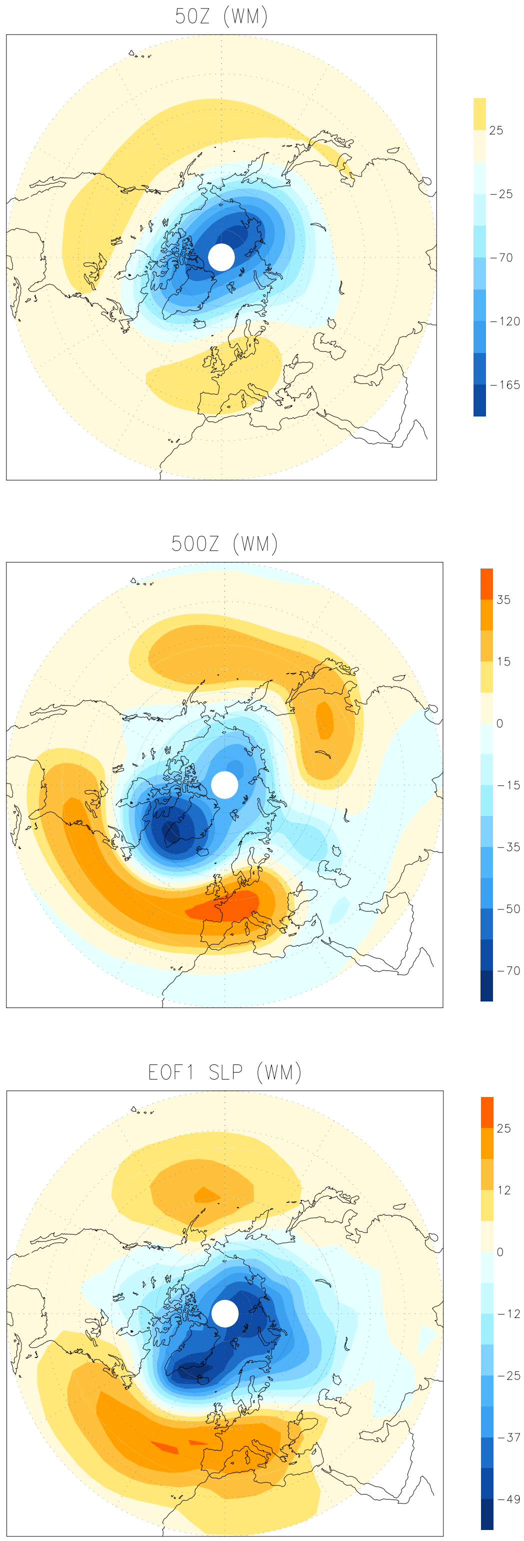

Figure 4. The leading EOF for SLP (expressed as Z_{1000}) and maps obtained by regressing geopotential height Z at 50-hPa and 500-hPa against the AO index. The figures show the change associated with one standard deviation of the AO index. The units are in metres. From Thompson and Wallace(1998), courtesy of David Thompson.

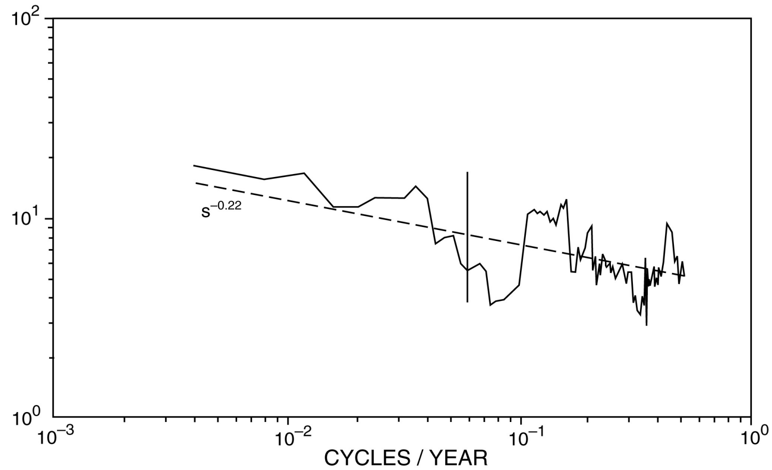

Figure 5. Multitaper spectral density estimate $\Phi(s)$ as a function of angular frequency s for the NAO index in Figure 1. From Wunsch(1999), courtesy of Carl Wunsch.

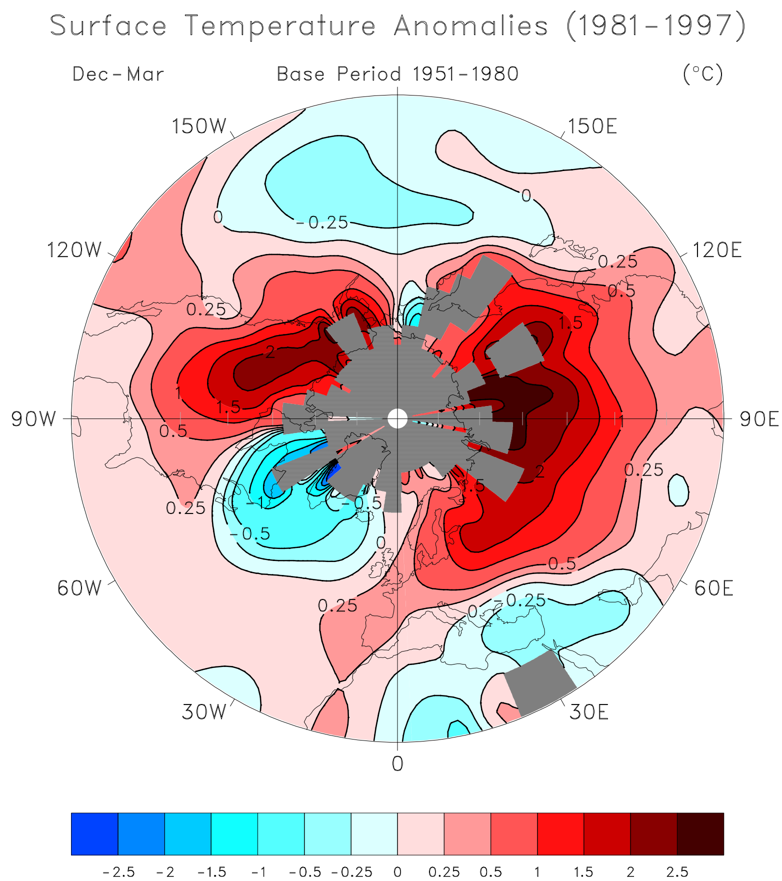

Figure 6. Average winter temperature 1981-1997 expressed as a departure from the 1951-1980 mean. From Hurrell(1996), courtesy of J. Hurrell.

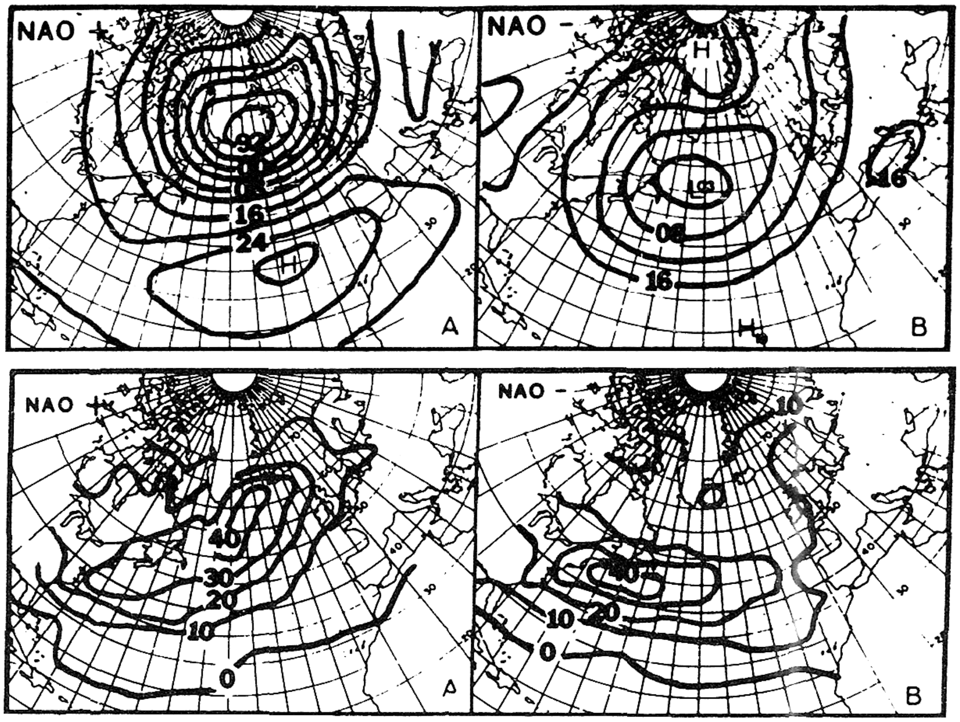

Figure 7. Top panels: The January mean SLP (hPa) for years between 1899 and 1986 with NAO index greater than or less than one standard deviation from the mean. Bottom panels: Isopleths of the frequency of wave cyclones per 5^o latitude x 5^o longitude tessarae for the nine winter months 1957-86 with the highest (left panel) and lowest (right panel) NAO index. From Rogers(1990).

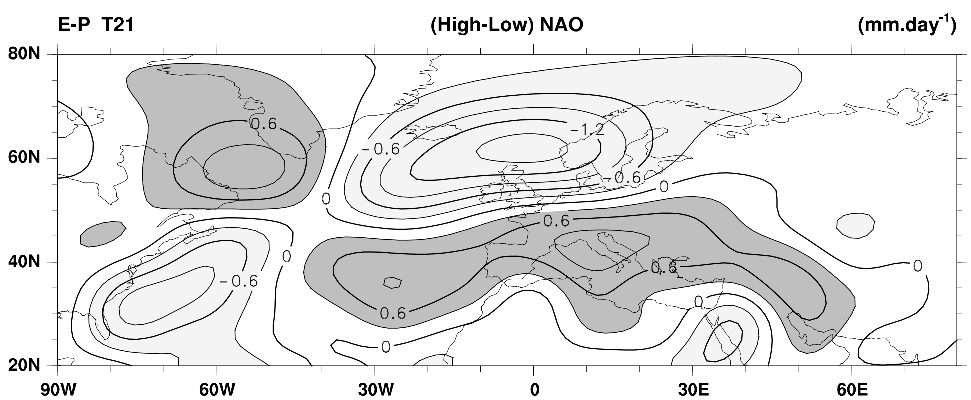

Figure 8. Evaporation minus precipitation for high NAO minus low NAO index, computed as the residual of the atmospheric moisture budget using ECMWF global analyses. See Hurrell(1995b) for details. Courtesy of J. Hurrell.

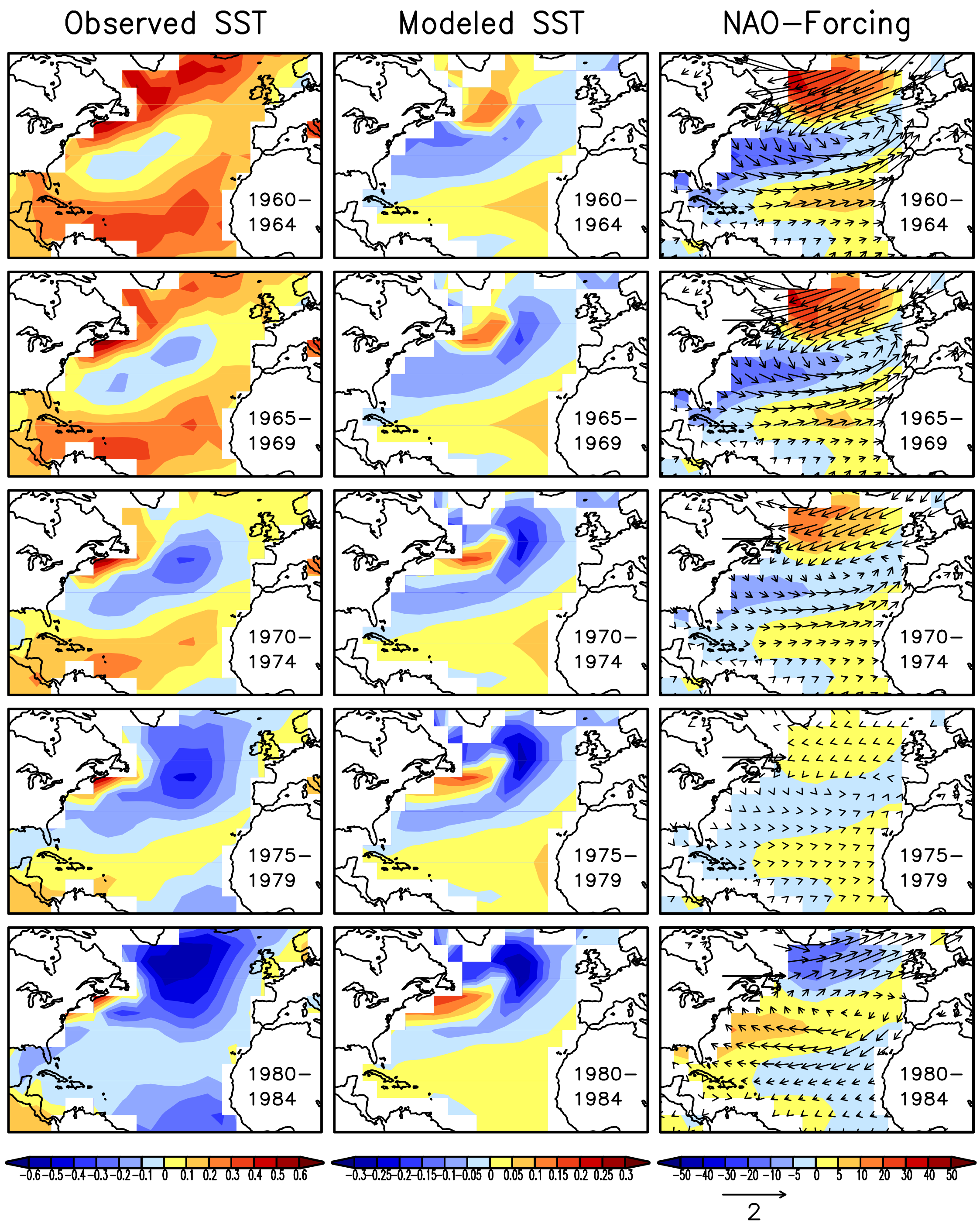

Figure 9. Interdecadal evolution of observed and modelled winter SST anomalies (^{o}C) for pentads from 1960-64 to 1980-84, and the NAO-related interdecadal heat flux (Wm^2) and near surface wind (ms^{-1}) fields that force the model. From Eden and Jung(2000), courtesy of Carsten Eden.

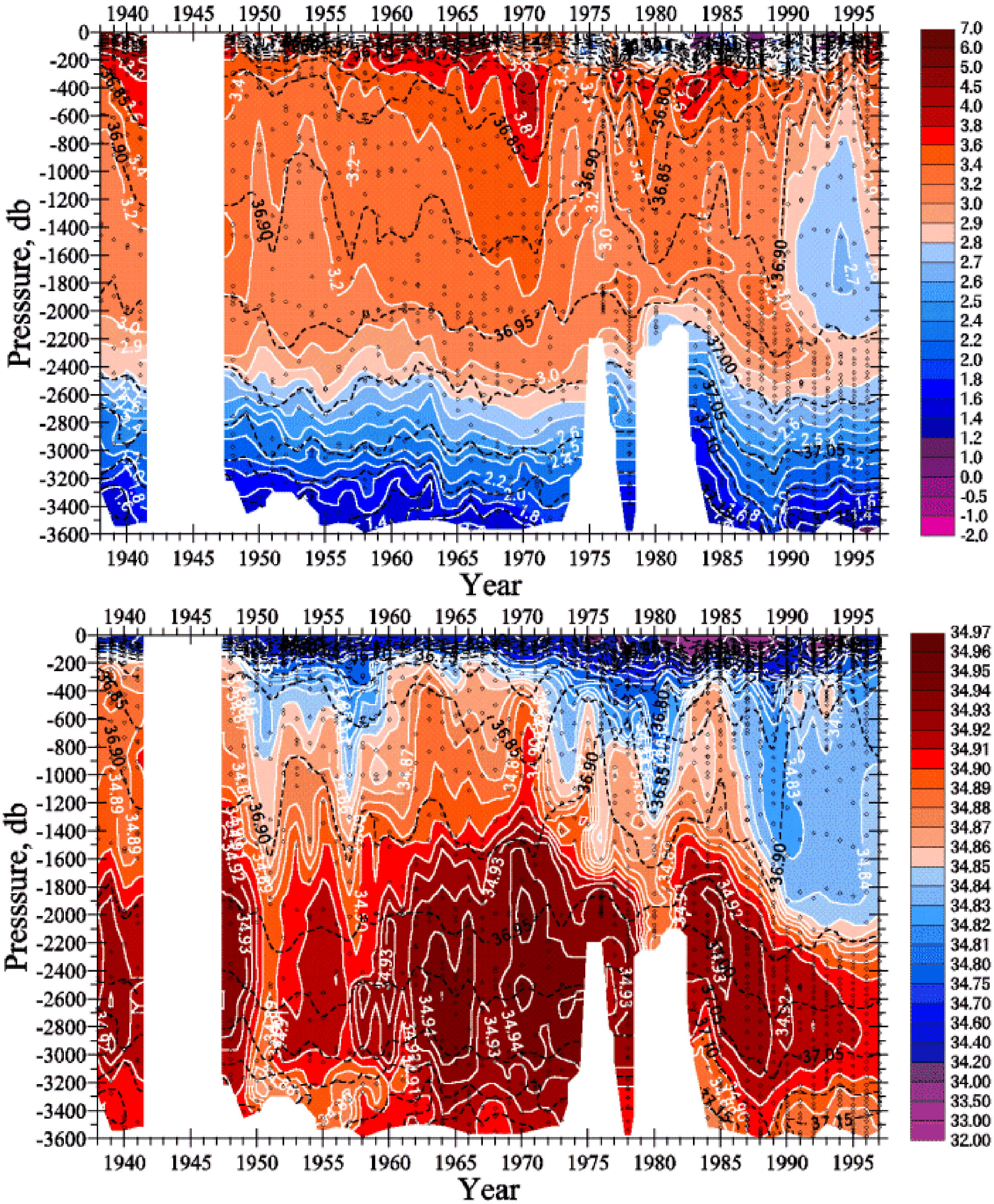

Figure 10. The time series of observed potential temperature (top panel) and salinity (bottom panel) in the Labrador Sea. Note the warming of the subsurface water in the 1960's when deep convection was suppressed, and the unprecedented cold and fresh water of the 1990's associated with the high NAO index of those years. Courtesy of Igor Yashayaev. (Extracted from http://dfomr.mar.dfo-mpo.gc.ca/science/ocean/woce/labsea/labsea_poster.html).

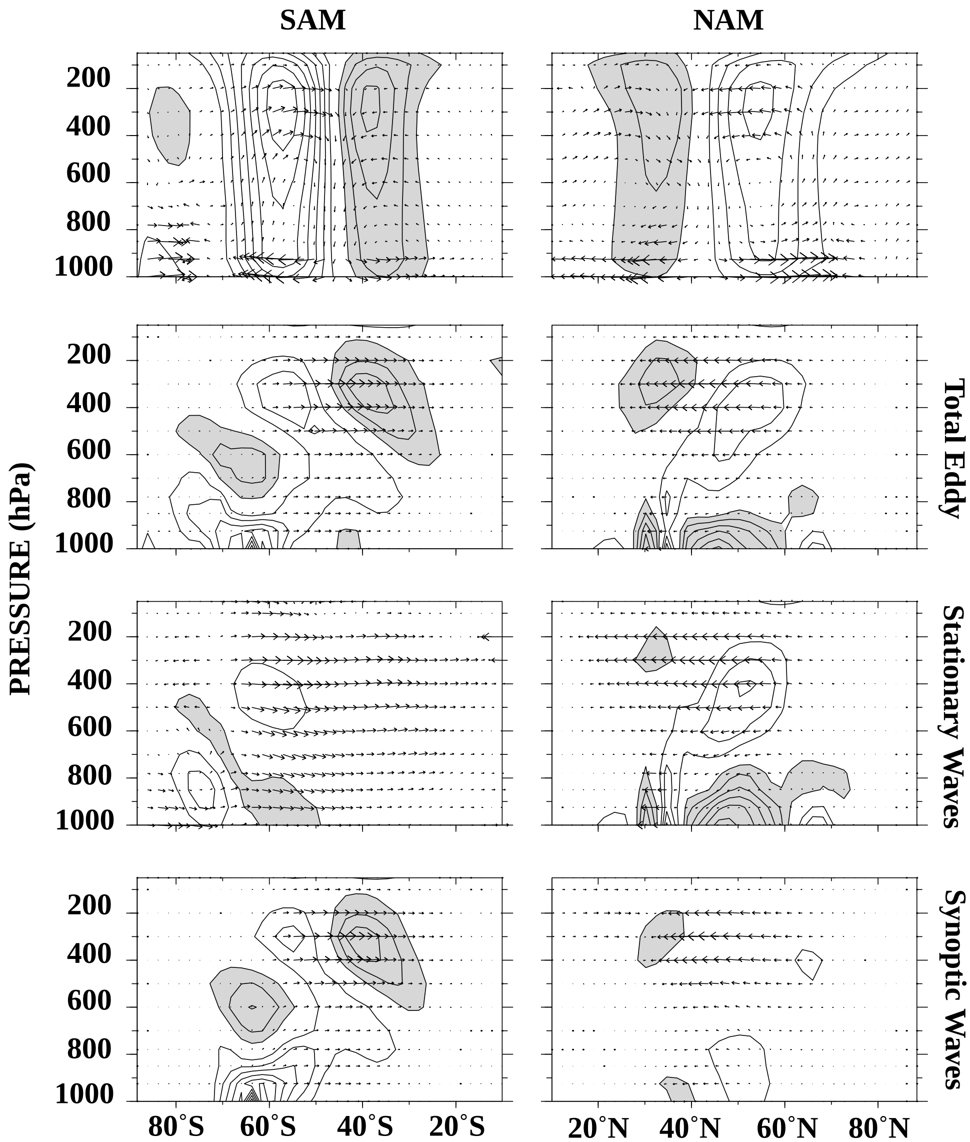

Figure 11. Top panels: Cross section of the zonal mean zonal wind regressed onto the SAM and NAM index (SAM/NAM is the Southern/Northern Angular Mode, that is the AAO and AO, respectively). The contour interval is 0.4 ms^{-1} per standard deviation of the index. Also shown as vectors is the corresponding zonal mean vertical and meridional winds. Other panels: Composite difference of the EP flux divergence contoured every 0.5 $\mathrm{ms}^{-1}\mathrm{day}^{-1}$. Easterly acceleration is shaded. Arrows indicate composite EP vector difference. All figures use output from the GFDL model. Taken from Limpasuvan and Hartmann(1999) where the details of the calculation can be found. Courtesy of V. Limpasuvan.

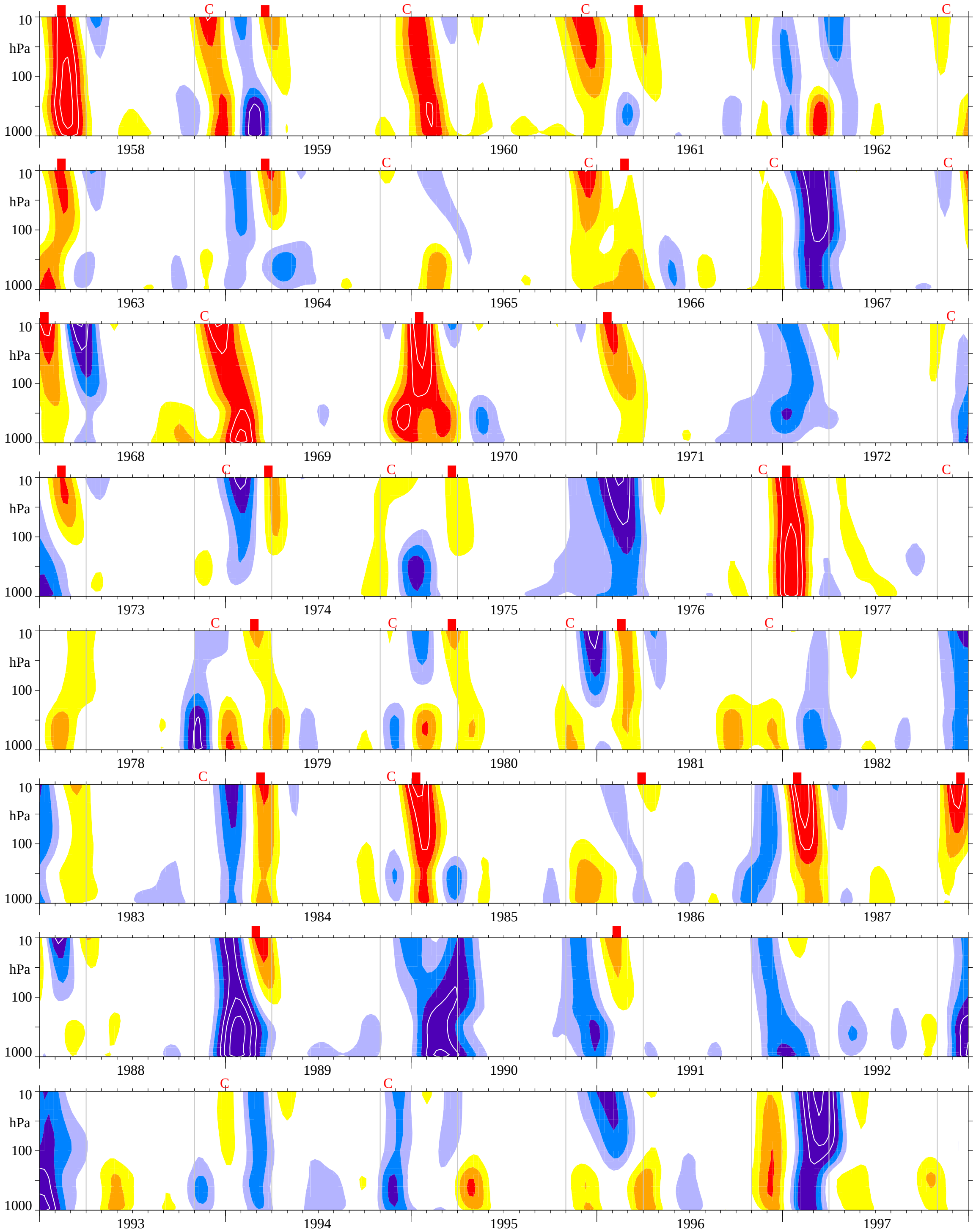

Figure 12. The evolution of the 90 day low-pass filtered projection of geopotential height on the AO computed from NCEP/NCAR reanalysis data. Red/blue shading indicates a weakened/strengthened circumpolar vortex, respectively, and red tick marks indicate major stratospheric warmings. From Baldwin and Dunkerton(1999) where the details of the calculation can be found. Note the tendency for events to propagate downward on a time scale of weeks. Courtesy of Mark Baldwin.

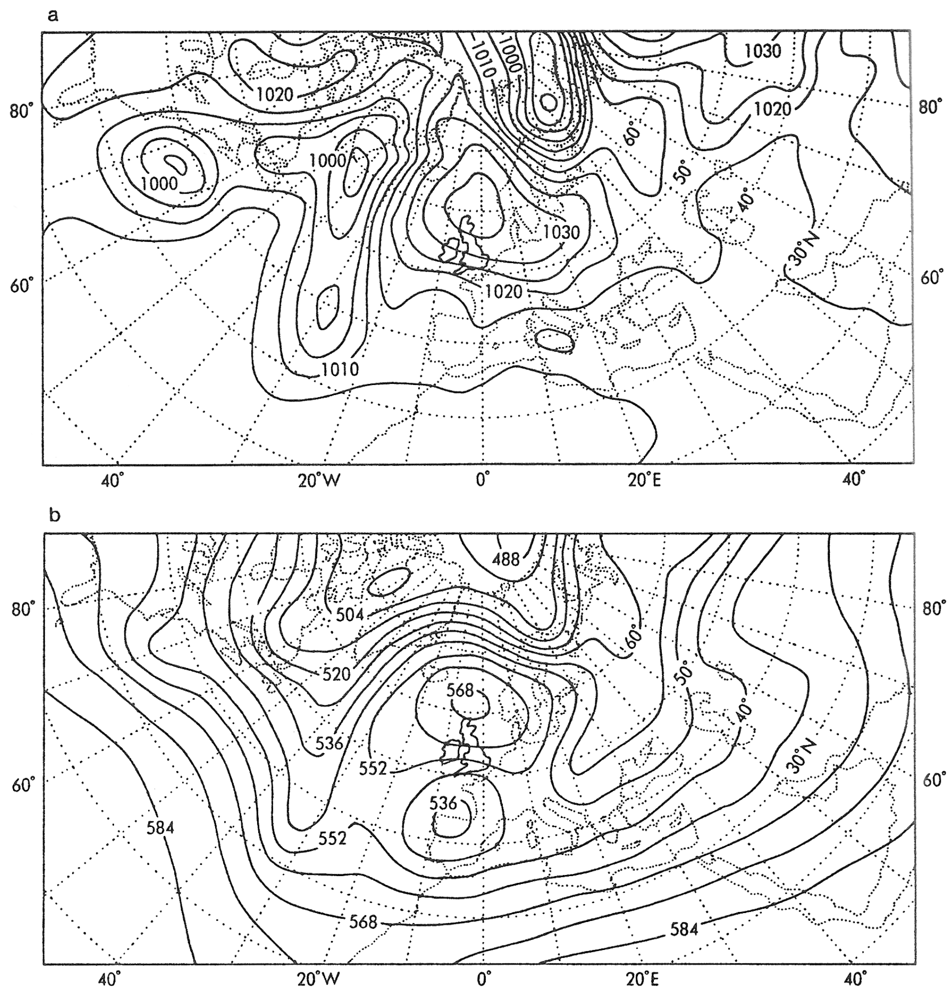

Figure 13. An example of a blocking pattern showing SLP (top panel) and 500 hPa height at 12Z 15 February 1983. From Shutts(1986).

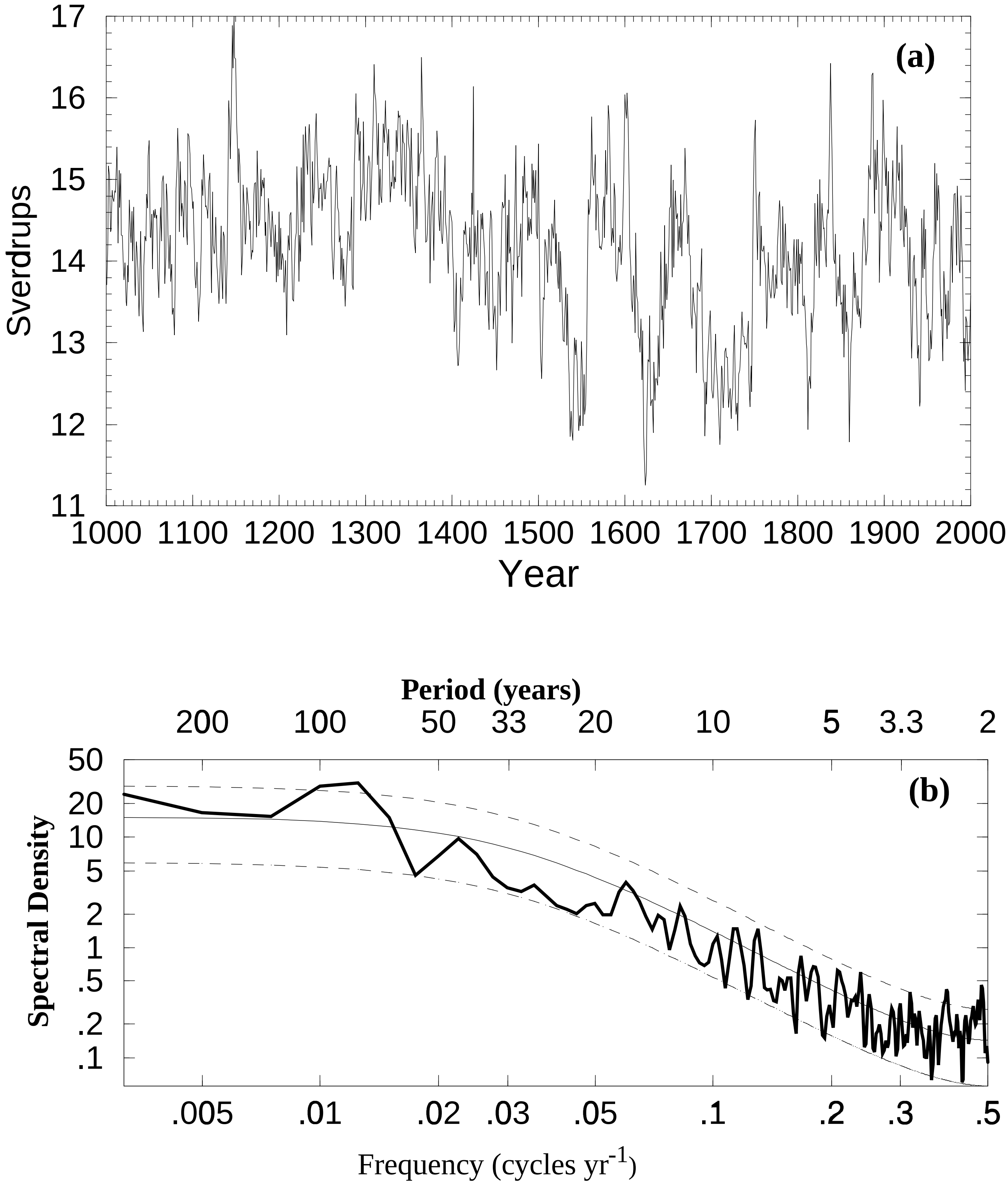

Figure 14. (a) Time series of the strength of the THC in the North Atlantic sector of the GFDL coupled model. (b) The Fourier spectrum of the time series in (a). From Delworth and Greatbatch(2000).

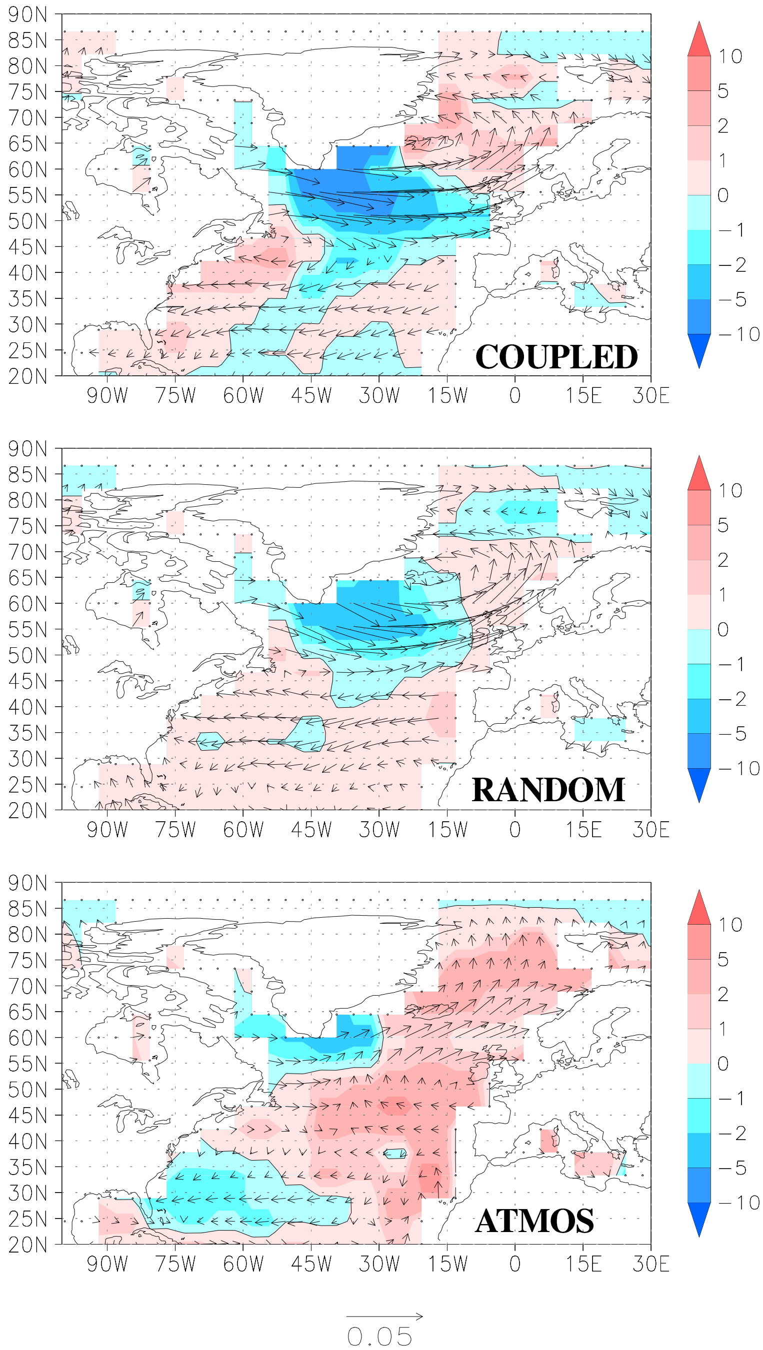

Figure 15. Regressions of surface heat flux (colour shaded in units of Wm^{-2}) and surface wind stress vectors (units are dyne $cm^{-2}Sv{-}1$, magnitude indicated by the arrow at the bottom) versus THC for a lag of -3 years. This denotes conditions 3 years prior to a maximum in the THC and corresponds approximately to a time when the flux anomalies have their largest amplitude. From Delworth and Greatbatch(2000).

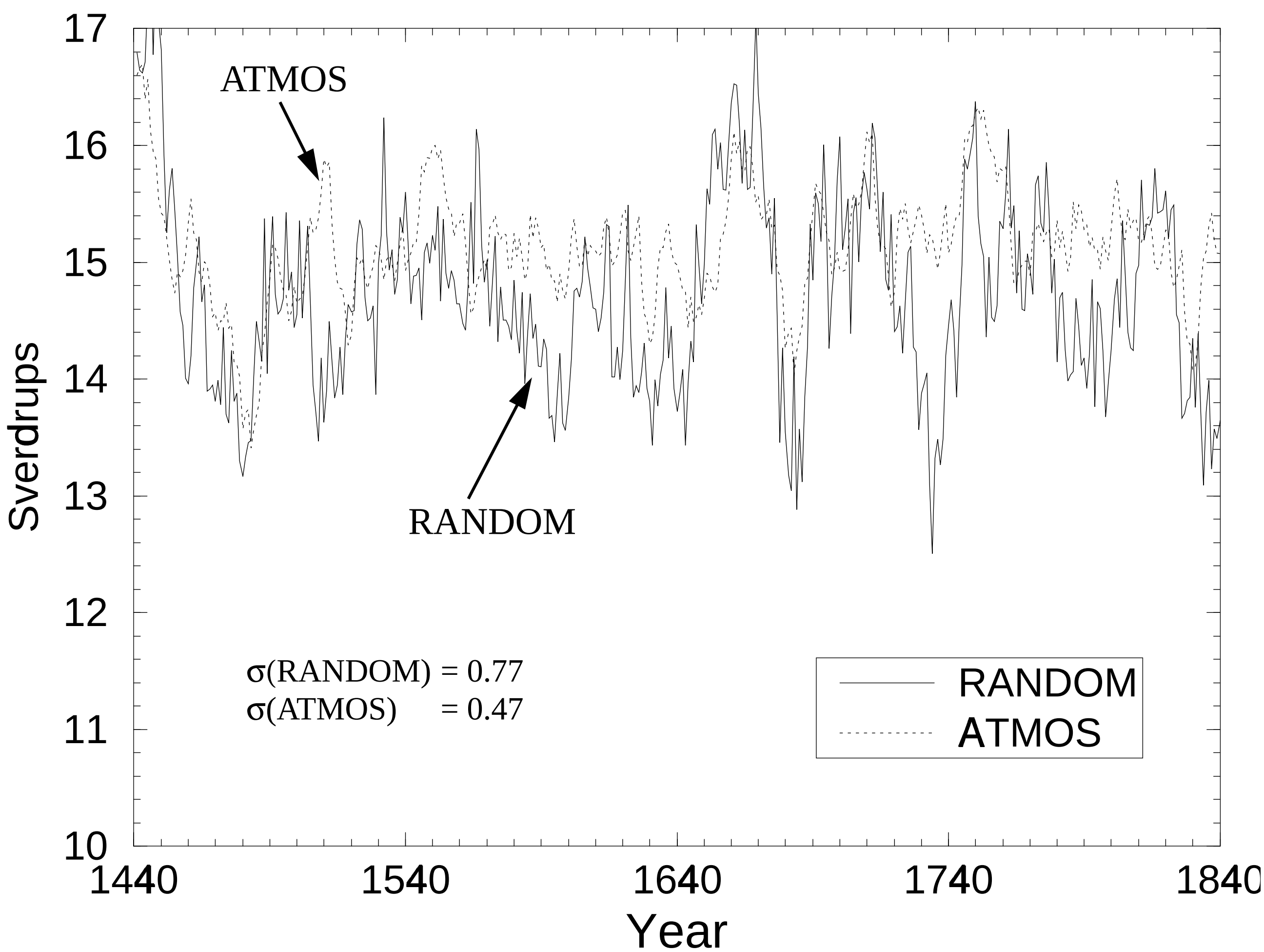

Figure 16. (a) Time series of the THC in RANDOM and ATMOS (see text for details). σ refers to the standard deviation of the variability in Sv. From Delworth and Greatbatch(2000).