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REVIEW

NORTH ATLANTIC CLIMATE VARIABILITY: PHENOMENA, IMPACTS AND MECHANISMS

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ABSTRACT

Variability of the North Atlantic Oscillation and the Tropical Atlantic dominate the climate of the North Atlantic sector, the underlying ocean and surrounding continents on interannual to decadal time scales. Here we review these phenomena, their climatic impacts and our present state of understanding of their underlying cause. Copyright © 2001 Royal Meteorological Society.

KEY WORDS: interannual time scale; North Atlantic Oscillation; climate variability

1. INTRODUCTION

Studies of North Atlantic climate variability have become a central focal point of climate research for the next decade. Scientists in both Europe and the US are planning co-ordinated observational, modelling and theoretical efforts focused on Atlantic Climate Variability as central elements to CLIVAR. We anticipate a major advancement of our current understanding of Atlantic Climate Variability, setting the stage for prototype predictability systems. In this article we review the nature, impacts and possible mechanisms of climate variability in the North Atlantic, and set out the framework that is being developed for advancing our understanding of their cause.

As described in Section 2, climate variability in the North Atlantic sector comprises three primary, but interrelated phenomena:

(I) Tropical Atlantic Variability (TAV):
a covarying fluctuation of tropical Atlantic sea surface temperature (SST) and trade winds straddling the Intertropical Convergence Zone (ITCZ)

(II) North Atlantic Oscillation (NAO):
a fluctuation in sea level pressure difference between the Icelandic Low and the Azores High; part of the Northern Hemisphere annular mode, the Arctic Oscillation (AO)

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(III) Atlantic Meridional Overturning Circulation (MOC):
fluctuations in the Atlantic's thermohaline circulation that may play a role in abrupt climate change.

In the tropics, low-frequency climate variability in countries surrounding the tropical Atlantic appears to be closely related to TAV through tropical Atlantic SST fluctuations. Variability in the rainfall of northeast Brazil correlates strongly with anomalies in the cross-equatorial SST gradient—see Section 3. Rainfall in subtropical West Africa also displays considerable dependence on the interhemispheric SST gradient. Furthermore, Atlantic equatorial SST anomalies appear to have a significant impact on anomalous rainfalls in the Guinea coastal region—see Table I.

In mid-latitudes, the leading mode of variability over the Atlantic region, the NAO, is profoundly linked to the leading mode of variability of the whole Northern Hemisphere circulation, the annular mode or Arctic Oscillation (AO). This suggests that Atlantic effects are more far-reaching and significant than previously thought. As we shall see, the NAO/AO has a climate significance that rivals the Pacific El Niño–Southern Oscillation (ENSO). The NAO exerts a dominant influence on temperatures, precipitation and storms, fisheries and ecosystems of the Atlantic sector and surrounding continents (see Table I). It is the major factor controlling air–sea interactions over the Atlantic and modulates the site and intensity of the sinking branch of the ocean's overturning circulation, the MOC. The NAO also seems to play a central role in real or perceived anthropogenic climate change. Understanding of the NAO and its time dependence appear central to three of the main questions in the global change debate: has the climate warmed, and if so why and how? In the North Atlantic, the MOC accounts for most of the oceanic heat transport and is a major player in maintaining the pole–equator temperature gradient. The possibility of a significant weakening of the MOC under global warming scenarios is a feature of coupled Global Climate Models (GCMs). This idea remains contentious. Yet because of its large potential impact, the possibility that increased fresh-water input and atmospheric high-latitude temperature anomalies could suppress the MOC must be taken seriously.

We argue here that the NAO, TAV and MOC are intimately connected on a wide range of time scales and should therefore be considered together. As described in Section 4, several new advances have been made in our understanding of the mechanisms that underlie these modes of variability. It appears that on interannual-to-decadal time scales, TAV and NAO may be interrelated. The NAO may provide an important extratropical forcing exciting tropical Atlantic variability. And the TAV may feedback on the NAO at interannual-to-decadal time scales by rearranging the Hadley Circulation. In mid-latitudes, shifts in the wind stress patterns and air–sea heat fluxes orchestrated by the NAO can lead to anomalous wind-driven gyres, thermohaline circulation and associated heat transport, thus impacting the MOC. Within the tropics, local air–sea feedback may be an important contributing factor in enhancing power in the cross-equatorial SST gradient and covarying pattern of trade winds, tropical SSTs and rainfall. On interannual time scales, the Pacific ENSO exerts an influence on the tropical western Atlantic and the Atlantic can influence the eastern equatorial Pacific too. Thus, 'remote' influences from the Pacific ENSO may also provide another major source of external forcing to excite Atlantic variability.

Finally, in Section 5 we discuss the prospects for predictability in the Atlantic sector and what needs to be done—both observationally and theoretically—to further our understanding and bring to fruition prototype predictability systems.

Before going on it should be emphasized that, to retain a manageable length, a number of important topics are not reviewed in depth. We have chosen to emphasize tropical/mid-latitude processes in the Northern Hemisphere. The link with the Arctic and atmosphere–ocean–ice interactions is not thoroughly reviewed, and much more could have been written on the role of the MOC in abrupt climate change. Nevertheless, we attempt to draw together in a comprehensible way many of the important strands and processes that play a role in the highly complex set of interactions that make up climate variability in the North Atlantic sector.

Table I. NAO-related climatic impacts that have been discussed in the literature. The first column provides a cross-reference number with Panel A of Plate 1. Other columns indicate the region of influence, the climatic variables affected and literature references, respectively. The reference list is representative but not exhaustive

No.*	Region of Influence	Phenomena	References
1	Arctic	Winter temperature, precipitation, sea ice distribution and movement.	Walsh and Johnson (1979); Kelly *et al.* (1982); Appenzeler *et al.* (1998); Kwok and Rothrock (1999); Parkinson *et al.* (1999).
2	North Atlantic	Winter storminess and ocean wave heights.	Dickson and Namias (1976); Carter and Draper (1988); Bacon and Carter (1991, 1993); Rogers (1997); Kushnir *et al.* (1997); The WASA Group (1998).
3	North Atlantic	Marine ecosystems, fisheries.	Dickson and Brander (1993); Post *et al.* (1997); Fromentin and Planque (1996); Dippner (1997); Fromentin *et al.* (1998); Reid *et al.* (1998); Planque and Taylor (1998); Marsh *et al.* (1999); Belgrano *et al.* (1999); Weyhenmeyer *et al.* (1999).
4	North Atlantic marginal seas	Atlantic sea ice anomalies.	Kelly *et al.* (1987); Walsh and Johnson (1979); Deser and Balckmon (1993); Deser *et al.* (2000).
5	Northwest Europe; Scandinavia	Winter temperature, rainfall and the severity of winters/effects on wildlife and plants.	Petterssen (1949); van Loon and Rogers (1978); Rogers (1985); Hurrell (1995b); Hurrell and van Loon (1997b); Thompson and Wallace (1998); Forchhammer *et al.* (1998); Loewe and Koslowski (1998); Wibig (1999); Chen and Hellstrom (1999); Post *et al.* (1999a,b); Post and Stenseth (1999).
6	Northeast Asia; Siberia	Winter temperature and rainfall.	Peng and Mysak (1993); Hurrell (1995b); Hurrell and van Loon (1997); Thompson and Wallace (1998); Livingstone (1999b).
7	British Isles	Rainfall, temperature and their ecological effects.	Rogers and van Loon (1979); Hurrell and van Loon (1997) Wibig (1999); Benner (1999); Kiely (1999); Milner *et al.* (1999); Colman and Davey (1999).
8	Central Europe	Winter temperature.	Walker and Bliss (1932); van Loon and Rogers (1978).
9	Southwestern Europe	Rainfall.	Rogers and van Loon (1979); Hurrell and van Loon (1997); Rodo *et al.* (1997); Ulbrich and Christoph (1999).
10	North America (Labrador, Hudson Bay, central US)	Winter temperature and snowfall/ effect on wildlife.	Hurrell (1995b); Hurrell (1996); Hartley and Keables (1998); Hartley, 1999; Post and Stenseth (1998); Post *et al.* (1999); Stenseth *et al.* (1999).
11	Mediterranean and the Middle East	Winter rainfall, temperature.	Metaxas *et al.* (1991); Kutiel and Kay, 1992; Cullen and de Menocal (2000).
12	North Africa—Morocco	Winter rainfall.	Lamb and Peppler (1987).
13	Central America, Caribbean	Rainfall.	Giannini *et al.* (2000).
14	Tropical North Atlantic (through SST)	Hurricane occurrence and tracks.	Landsea *et al.* (1999); Shapiro and Goldenberg (1998).

* The numbers in this column refer to those in the grey circles that appear in Plate 1.

2. THE PHENOMENOLOGY OF ATLANTIC CLIMATE VARIABILITY

2.1. The North Atlantic Oscillation

More than two centuries ago missionaries noticed that year-to-year fluctuation in wintertime air temperatures on both sides of Iceland were often out of phase with one another—see van Loon and Rogers (1978) for historical references. When temperatures are below normal over Greenland, they are above normal in Scandinavia, and vice versa. Simultaneously, coherent fluctuations in temperatures, rainfall and sea level pressure were documented, reaching eastwards to central Europe, southwards to subtropical West Africa and westwards to North America. This mode of climate variability is now known as the North Atlantic Oscillation (NAO), a name first cited by Walker (1924; see also Rogers, 1984). It is now firmly established that its fluctuations influence climate from North America to Siberia and from the Arctic Ocean to the equator and perhaps beyond.

2.1.1. Spatial pattern. The NAO (Figure 1) can be best discerned when time-averaged (monthly or, preferably, seasonal) atmospheric fields are analysed (Esbensen, 1984; Kushnir and Wallace, 1989), although it is also evident in daily data. Walker and Bliss (1932) described the NAO as 'expressing the tendency for pressure to be low near Iceland in winter when it is high near the Azores and south west Europe' and vice versa. Indeed its manifestation at sea level is a dipole roughly overlapping the Icelandic Low and the Azores High (see, e.g. van Loon and Rogers, 1978; Hurrell, 1995a). At different phases of the NAO, the strength and direction of the storm tracks shift, the intensity and path of storms being modulated by the time-averaged flow (Rogers, 1990).

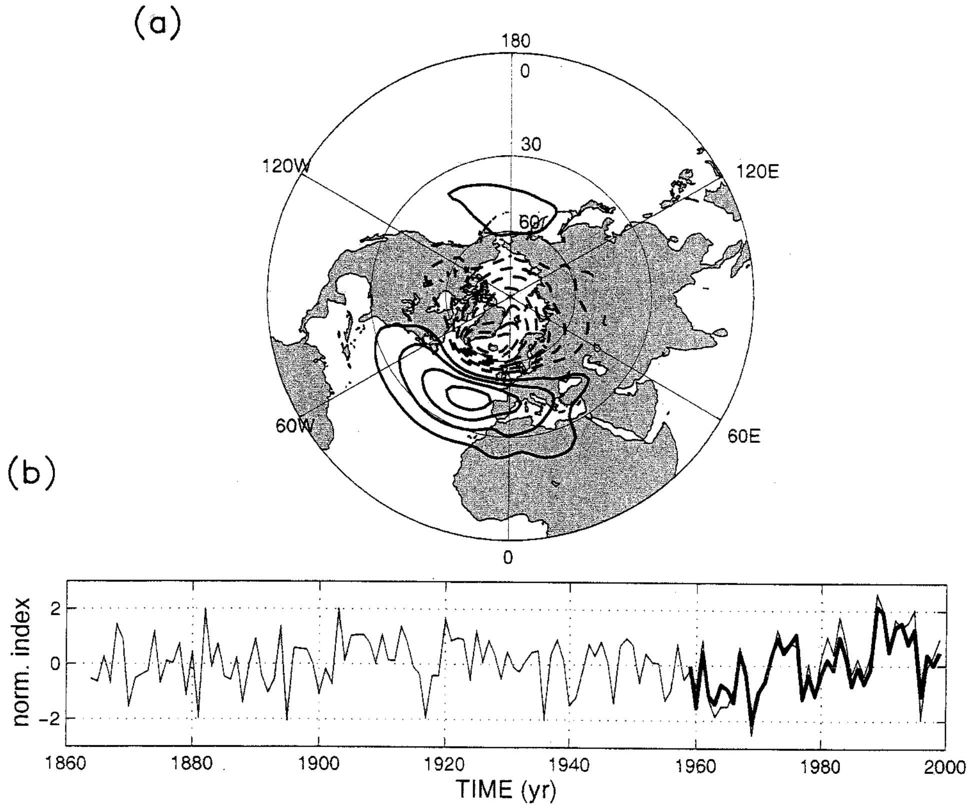


Figure 1. (a) Regression map of Northern Hemisphere SLP anomalies in winter (December–March 1958–1998) onto the first principal component of SLP anomalies over the North Atlantic sector (20°–70°N/100°W–20°E). (b) Time series of Hurrell's NAO index (thin curve) and the first principal component of SLP (thick curve). Both time series are normalized by their standard deviation. The SLP data are taken from the NCEP–NCAR reanalysis

The NAO appears to be part of a hemispheric mode of variability. Thompson and Wallace (1998) and Thompson *et al.* (2000a,b) discuss the concept of the 'Arctic Oscillation' (AO), or Northern Hemisphere 'annular mode', which they defined as the leading empirical orthogonal function (EOF) of wintertime monthly mean Northern Hemisphere sea level pressure (SLP). Wallace (2000) argues that the NAO and AO represent a single phenomenon viewed through two paradigms (but see the discussions in Deser, 2000 and Ambaum *et al.*, 2001). The NAO may be viewed as the regional expression of the AO. The AO may be thought of as the leading wintertime hemispheric low-frequency mode of variability of SLP, while the NAO is the leading mode in the Atlantic basin. Maps of the two modes are nearly indistinguishable (except for the Pacific region). In this review, where the emphasis is on Atlantic Climate Variability, we use NAO unless a distinction needs to be made.

In contrast to the climatological mean flow over the North Atlantic, the NAO/AO has a pronounced 'equivalent barotropic' structure (vertical phase lines; Wallace and Gutzler, 1981; Kushnir and Wallace, 1989) and increases in amplitude with height in rough proportion to the strength of the mean zonal wind, coupling the troposphere with the stratosphere (Perlwitz and Graf, 1995).

Figure 2(a) shows the regression of SST on the NAO index during winter. It reveals a tri-polar pattern—a cold subpolar region, warmth in mid-latitudes and a cold region between the equator and 30°N. This is also the leading pattern of SST variability during winter. Its emergence is consistent with the spatial form of the anomalous surface fluxes associated with the NAO pattern, as pointed out by Cayan (1992a,b); see Figure 2(b). It appears that the strength of the correlation between the NAO and SST increases when the NAO index leads SST, indicating that SST is responding to atmospheric forcing on monthly to seasonal time scales (Frankignoul, 1985; Battisti *et al.*, 1995; Delworth, 1996; Hall and Manabe, 1997).

2.1.2. Temporal characteristics. The NAO is most pronounced in amplitude and areal coverage during winter (December–March) when it accounts for some 37% of the monthly time series of December, January, February (DJF) 500 hPa height variability over the Atlantic (Wallace and Gutzler, 1981; Esbensen, 1984; Barnston and Livezey, 1987; Kushnir and Wallace, 1989; Wallace, 1996; Cayan 1992a). The NAO is the only mode found in all seasons (Barnston and Livezey, 1987; Clinet and Martin, 1992) and Rogers (1990) showed that it accounts for the largest amount of interannual variability in monthly North Atlantic SLP in all but 4 months of the year. The second mode of variability—the East Atlantic (EA) pattern—explains some 19% of monthly DJF 500 hPa variability over the Atlantic. Feldstein (2000) finds that the NAO and PNA explain roughly equivalent fractions—18%—of the seasonal *hemispheric* 300 hPa height field variability.

The NAO is often defined by an index of normalized, time-averaged pressure differences between stations representing its two centres of action, such as the Azores and Iceland (Rogers, 1984; Hurrell, 1995a). Such a simple index clearly cannot take account of the possibility that the centres of the actual pattern may not overlap with these locations, nor can it accurately capture the seasonal variations in the NAO (Barnston and Livezey, 1987). However, there is a key advantage to the use of such an index because the existing weather records allow it to be extended back in time to at least 1864. When the index is correlated or regressed with gridded surface pressure data, the resulting north–south dipole pattern defines the spatial pattern of the NAO.

The winter NAO index, defined as the difference between the normalized mean SLP in the Azores and Iceland, is shown in Figure 1(b) since 1864. Positive values indicate stronger-than-usual westerly winds. It is evident from Figure 1(b) that the NAO index has exhibited considerable variability over the past 100 years. From the turn of the century until about 1930 (with the exception of the 1916–1919 winters), the NAO was high and so stronger-than-usual winds carried the moderating influence of the ocean over Europe contributing to the higher-than-normal European temperatures during this period (e.g. Rogers, 1985). From the early 1940s until the early 1970s, the NAO index exhibited a downward trend, and corresponded to a period in which European wintertime temperatures were frequently lower than normal (van Loon and Williams, 1976; see also Plate 1a). A sharp increase has occurred over the past 25 years. Since 1980, the NAO has remained in a strongly positive phase, and displayed an upward trend (but may

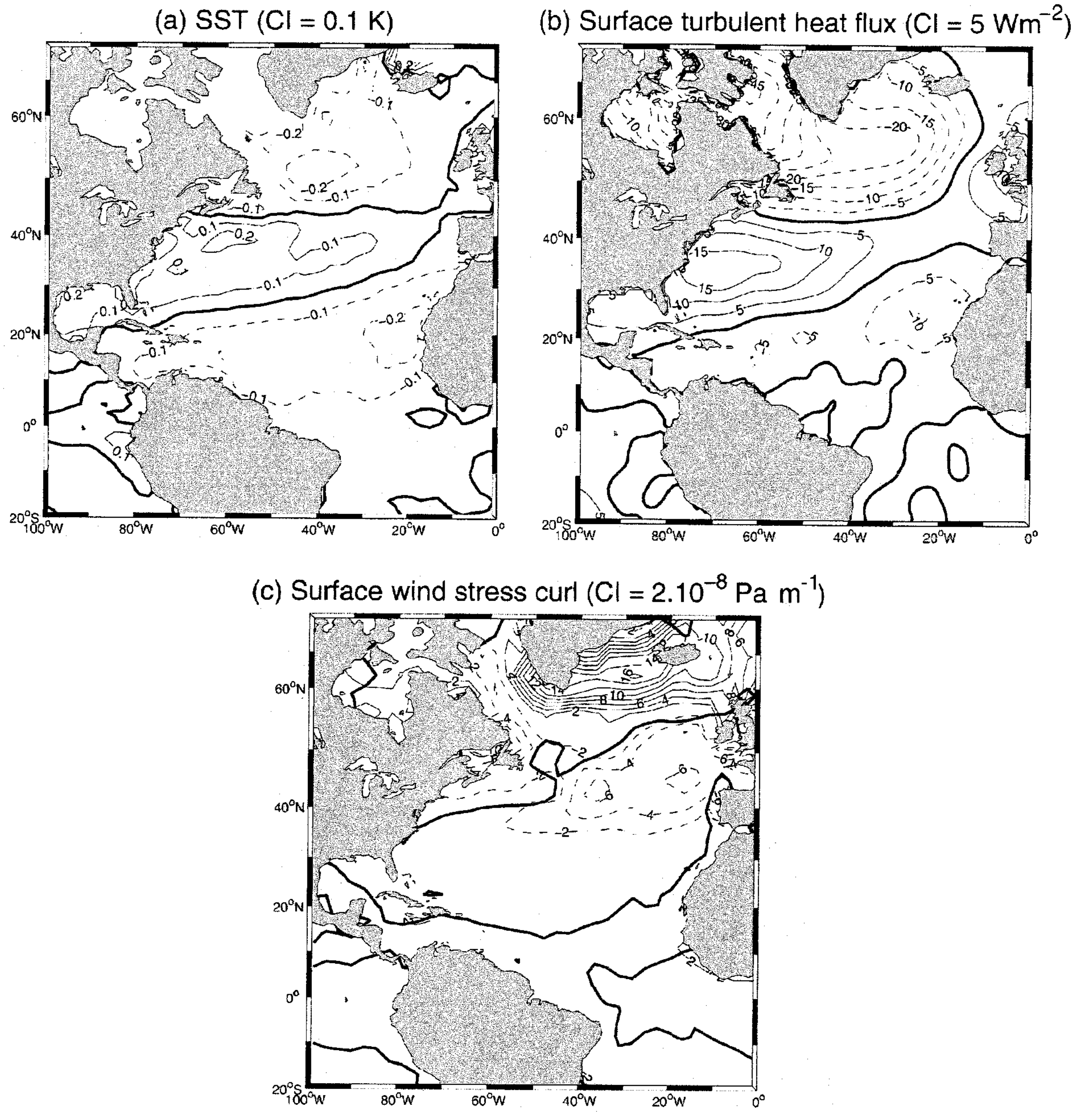


Figure 2. Regression maps of (a) SST (CI = 0.1 K, negative dashed), (b) surface turbulent heat flux (latent + sensible, CI = 5 Wm^{-2}, dashed out of the ocean) and (c) surface wind stress curl (CI = 2×10^{-3} Pa m^{-1}, dashed for anticyclonic) anomalies onto the NAO index shown in Figure 1(b). The thick black line is the zero wind curl line of the climatological winds. The SST and wind stress curl data come from the NCEP–NCAR reanalysis, while the surface heat flux anomalies come from Da Silva *et al.* (1994). A linear trend was removed to all datasets prior to computing the linear regressions

now be abating), so that at the turn of the 20th century winters have exhibited the most pronounced positive indices ever recorded (with the notable exception of 1996). This situation contributed much to the observed warming in Northern Hemisphere surface temperatures over the past two decades (Hurrell, 1995a, 1996; Wallace *et al.*, 1995; and Section 3). The increased AO during the past 30 years (Thompson and Wallace, 1998) also corresponds to a stronger polar vortex, coinciding with a depletion of stratospheric ozone levels above the polar cap.

The spectrum of the wintertime NAO index (Figure 3) is slightly red, with power increasing with period. As discussed in Wunsch (1999) and Stephenson *et al.* (2000)—see also Deser and Blackmon (1993)—the index shows enhanced power in some frequency bands, although not large enough to be statistically significant. The slight maximum with a period between 2 and 3 years is most noticeable in Northern Hemisphere mid-latitude SST, although the mechanism behind this peak remains uncertain (Stephenson *et al.*, 2000 who discuss the 2–3 year variability in the Northern Hemisphere SLP). When the NAO index is taken as the first principal component of SLP anomalies over the Atlantic sector (Figure 3, thick curve), similar spectral features are found. Note, however, that principal component analysis, by taking into account many more observations, reduces the noise inherent in station land data. At high frequencies, its power spectrum has a larger slope than that based on Hurrell's index.

Paleoclimate indicators, such as tree rings and ocean sediments, have been used as proxies for the NAO and can be extended back several hundred years. These suggest that the NAO is robust and has remained coherent for at least 1000 years and has been coherent over long periods of the 700-year stable isotope record from the GISP-2 ice core from central Greenland (Barlow *et al.*, 1993; White *et al.*, 1997). Similar evidence emerges from the tree ring data (Cook *et al.*, 1998). Wanner (personal communication, 1998)

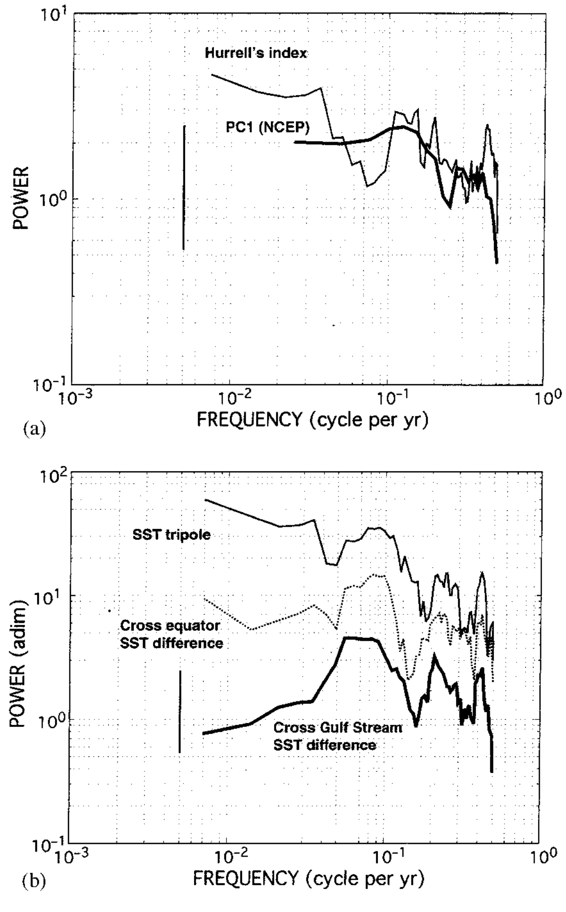


Figure 3. Power spectra of (linearly detrended) SLP and SST indices, estimated using the multi-taper method. (a) NAO index from Hurrell (thin curve) and from a principal component analysis of winter SLP anomalies (December–March, 1958–1998) over the North Atlantic sector (20°–70°N/100°W–20°E) from the NCEP–NCAR reanalysis (thick curve). (b) Indices of cross-Gulf Stream ΔT_{GS} (black continuous line), cross-equatorial ΔT_{EQ} (black dashed line) and tripole SST anomalies (grey line). The indices are ΔT_{GS} the difference between SST averaged over [40°–55°N/60°–40°W and 25°–35°N/80°–60°W] (late winter-February–April), and ΔT_{EQ} [5°–20°N/40°–20°W and 15°–5°S/15°W–5°E] (all calendar months). The SST tripole index is defined as the first principal component of winter (December–March) SST anomalies over the North Atlantic (20°–70°N). All SST data are from Kaplan *et al.* (1998). For all spectra, the 95% contour interval is shown by the vertical line, and the number of tapers has been set to $K = 7$

reconstructs the NAO from proxy and documentary sources and hypothesizes that decadal-scale cold relapses during the Late Maunder Minimum (1675–1715) were related to successive negative NAO events.

2.1.3. Covarying patterns in SST and the interior ocean. There are clear indications that the North Atlantic ocean varies significantly with the NAO above. The leading pattern of SST variability in the North Atlantic—the tripole plotted in Figure 2(a)—is a direct response of the ocean to the anomalous air–sea fluxes derived by the NAO shown in Figure 2(b). Because the NAO is most active from December–March, the SST tripole is most energetic in late winter (it needs roughly 1 month to integrate the NAO forcing). As shown in Watanabe and Kimoto (2000a), the tripole persists longer (for about a year) than would be expected from local damping due to air–sea interaction (which yields a decay scale of ~ 3 months—see Frankignoul *et al.*, 1998). This is related in part to the formation of the upper seasonal thermocline in summer, which reduces the damping of SST anomalies formed during the previous winter, and to their subsequent re-entrainment into the mixed layer the following winter (the so-called re-emergence mechanism discussed in Section 4.1.2—see, e.g. Alexander and Deser, 1995). The tripole pattern has a red spectrum—see Figure 3(b)—that is broadly consistent with the statistical model of Hasselman (1976), which invokes the thermal inertia of the mixed layer as a reddening mechanism. One notes, however, a hint of increased power in the decadal band (but at a slightly longer period than that seen in the NAO index, see Figure 3(a)) and increased power at the longest time scales, as was found in Hurrell's NAO index.

Deser and Blackmon (1993) present spectra of EOFs of SST, marine air temperature, SLP and surface winds that show an increase in variance near the decadal band. Indeed, elements of the tripole show pronounced decadal signatures. Czaja and Marshall (2001), using a long historical record of SST data, define an SST index, $\Delta T_{GS} = T_N - T_S$, the difference between the SST in boxes to the north and south of the separated Gulf Stream—this index is closely associated with the 'Dipole mode' of Deser and Blackmon (1993). It is here in the vicinity of strong western boundary currents and recirculation that we might expect to see a signature of ocean circulation imprint itself on SST. Indeed, the spectrum of ΔT_{GS} shown in Figure 3 reveals a broad peak in the 10–20-year band, with a marked reduction in power at low frequencies. Czaja and Marshall (2001) study the temporal behaviour of ΔT_{GS}—it does not simply die away after a couple of years, but reappears with an opposite sign, roughly 6 years after it has been generated. They argue that such behaviour is likely to be a consequence of advection of heat by anomalous ocean currents and possible feedback on the overlying atmosphere. Spreading of SST anomalies along the path of the Gulf Stream and North Atlantic Current (see Figure 4, modified from Sutton and Allen, 1997—see also the observations of Hansen and Bezdek, 1996 and the modelling study of Visbeck *et al.*, 1998 and Krahmann *et al.*, 2001). Winter SST anomalies born in the western subtropical gyre appear to propagate eastwards with a transit time of a decade or so.

Subsurface ocean observations provide an even clearer depiction of long-term climate variability because the effect of the annual cycle and month-to-month variability in the atmospheric circulation decays rapidly with depth—see Levitus (1989) and Houghton (1996). Variability in the temperature of the upper subtropical ocean has also been linked with fluctuations in the NAO index (Molinari *et al.*, 1997). Curry and McCartney (2001) discuss observed interannual to interdecadal variations in the intensity of ocean gyres and relate them to NAO forcings. A large freshening event (dubbed the Great Salinity Anomaly; GSA) occurred in the subpolar gyre in the early 1970s (Dickson *et al.*, 1988; Levitus, 1989). Observations suggest it may have been linked to coherent changes in the overlying atmospheric circulation and an extreme manifestation of quasi-decadal fluctuations (Kushnir, 1994; Hansen and Bezdek, 1996; Reverdin *et al.*, 1997; Belkin *et al.*, 1998).

The intensity of ocean convection is also observed to covary with the NAO, particularly in the Labrador Sea—see Dickson *et al.* (1996) and Lab Sea Group (1998). The link between propagating SST anomalies and the properties of water mixed at the end of winter (McCartney *et al.*, 1997) shows that the end-product of the transformation process, Labrador Sea Water (LSW), underwent extended warming and cooling trends consistent with the phase of the NAO; Figure 5. There is also evidence that hydrographic changes of decadal scale in the Arctic and subarctic seas are able to feed south across the

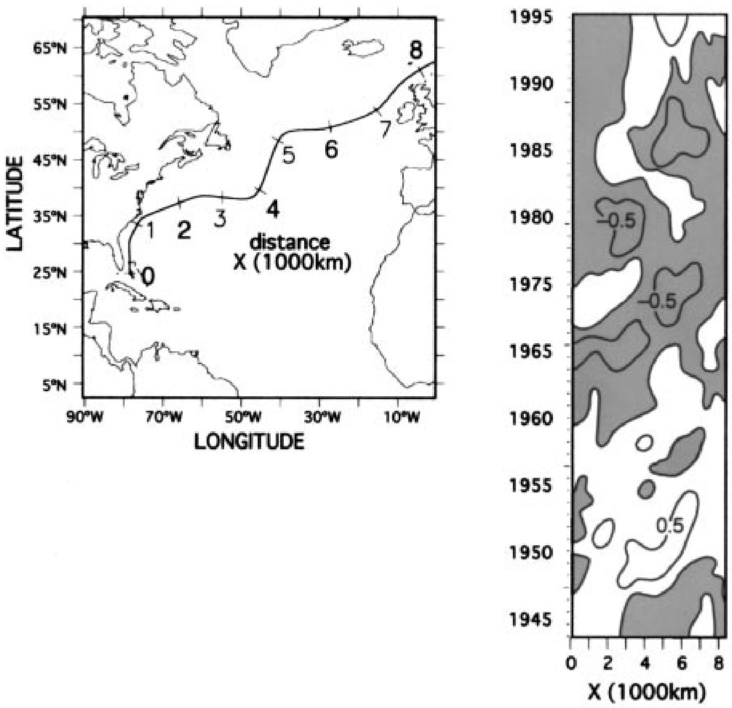


Figure 4. Left panel: pathways of interannual propagation of SST anomalies along the North Atlantic Current. Sutton and Allen (1997) find that it takes 12–14 years to travel from 0 to 7500 km, yielding an average speed of 1.7 cm s^{-1}. Right panel: Hovmoller diagram of winter SST anomalies as a function of time and position along the pathway marked on the left panel. The contour interval is 0.5°C and negative values are shaded. Modified from Sutton and Allen (1997), reproduced with permission of Nature

deep northern overflows to cause hydrographic changes in the deep and abyssal layers of the Labrador Sea—see Dickson *et al.* (2001). These variations appear to be large and long sustained.

2.2. Tropical Atlantic variability

The dominant low-frequency climate phenomenon in the tropical Atlantic is the covarying fluctuation of tropical SST and trade winds (Nobre and Shukla, 1996; and Figure 6(a)). These fluctuations exhibit a pattern of basin-wide SST anomalies straddling the mean position of the ITCZ, with concomitant changes in trade wind intensity. The Northern Hemisphere trades, and the associated cross-equatorial flow, seem to depend sensitively on SST. Weaker-than-normal trades in the northern subtropics are associated with positive SST anomalies beneath them, and vice versa (Nobre and Shukla, 1996; Enfield and Mayer, 1997). The northern trades and location of the ITCZ are also affected by SST changes south of the equator, such that colder-than-normal SST in the South Atlantic correspond to weaker-than-normal trades in the North Atlantic and a smaller-than-normal southward displacement of the ITCZ during the boreal spring. Early statistical analysis of SST data revealed a dipolar structure and led investigators to dub the pattern the 'tropical Atlantic SST dipole'. However, SST variability north and south of the equator are not correlated (Houghton and Tourre, 1992; Enfield and Mayer, 1997; Mehta, 1998; Rajagopalan *et al.*, 1998), suggesting that it is the change in the cross-ITCZ SST contrast to which the trades respond. While it is now clear that SST variability across the Atlantic ITCZ is not well described as a temporally coherent seesaw fluctuation, climate variability in the tropical Atlantic is clearly sensitive to cross-equatorial SST differences (Hastenrath and Heller, 1977; Folland *et al.*, 1991; Servain, 1991; Ward and Folland, 1991).

SST fluctuations on both sides of the equator affect the interhemispheric SST gradient, ΔT_{EQ}. The southern-most lobe of the SST tripole—see Figure 2(a)—is under the influence of the NAO reaching down into the subtropics/tropics. The build-up of the tropical/subtropical North Atlantic SST anomaly

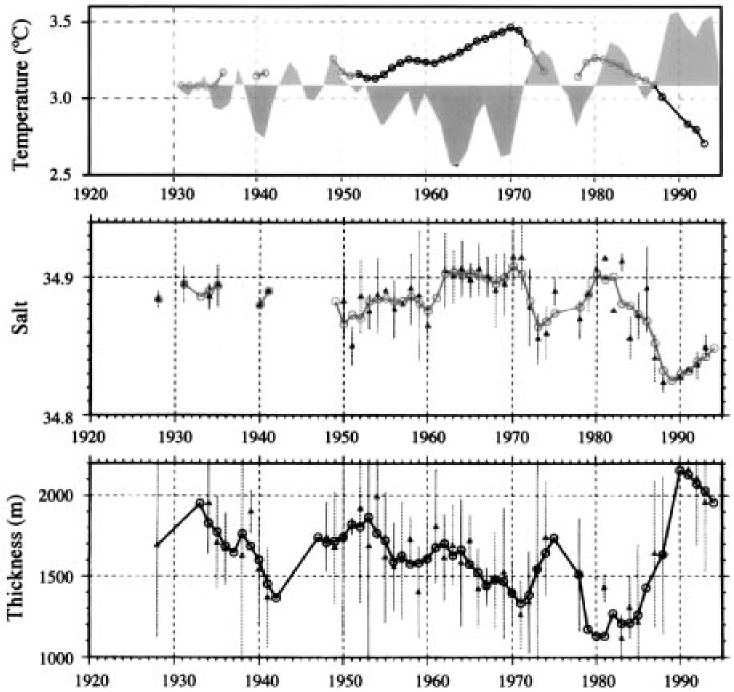


Figure 5. Time history of (top) the temperature of LSW, plotted along with the NAO index-shaded (middle) salinity in the Labrador Sea and (bottom) thickness of LSW core. (Figure from p. 21 from McCartney MS, Curry RG, Bezdex HF. 1997. North Atlantic's transformation pipeline chills and redistributes subtropical water. *Oceanus* **39**: 19–23. Reproduced by permission of WHOI.) As the NAO has strengthened, LSW has become colder, fresher and more voluminous. We also observe much interannual variability superimposed on the trend

lags behind the local change in the wind circulation by a month or two (Nobre and Shukla, 1996), slightly shorter than the lag between the atmosphere and ocean in mid-latitudes (Frankignoul, 1985). As the SST anomaly builds up, the Northern Hemisphere trades adjust. Considerable SST fluctuations also occur in the subtropics of the Southern Hemisphere (Venegas *et al.*, 1997). Their nature is not entirely resolved, but they appear to be linked to the interannual variability in the South Atlantic subtropical high.

The time series of the joint SST and wind pattern (Figure 6(b)) clearly contain long-term components and noticeable imprints of decadal variability. The spectrum of the cross-equatorial SST gradient, ΔT_{EQ}, shows somewhat enhanced power on time scales of 10–20 years—see Figure 3—but not markedly so.

A mode of variability similar to the Pacific ENSO has also been identified in the tropical Atlantic Ocean (Covey and Hastenrath, 1978; Philander, 1986; Zebiak, 1993; Carton and Huang, 1994). Although they are much weaker than their Pacific counterparts, the Atlantic equatorial SST anomalies can, for example, have an effect on rainfall in the Gulf of Guinea (Wagner and da Silva, 1994). As in the Pacific, equatorial waves and remote wind forcing play a significant role in the generation of SST anomalies at the interannual time scale (Hirst and Hastenrath, 1983; McCreary *et al.*, 1984). These perturbations in equatorial winds and SST are not as dominant as their Pacific counterparts, apparently because the Atlantic basin is much narrower than the Pacific, and do not appear to be consistently sustained (Zebiak, 1993). However, they do exert some influence over the adjacent landmasses.

2.2.1. The relation between NAO and TAV. There exists a strong link between tropical climate variability and the NAO. Changes in trade winds, governed by fluctuations in the strength and location of the Azores High, impact SST beneath them through associated surface heat exchanges and entrainment at the bottom of the ocean mixed layer. Thus the SST tripole and interhemispheric SST gradients share common SST anomalies between the equator and $\sim 30°N$. Consistent with the NAO impacting the northern subtropical Atlantic, the interannual variability of SST shown in Figure 6(a) is strongest in March–May (Nobre and Shukla, 1996), lagging by 1–2 months the most active NAO season (January–March).

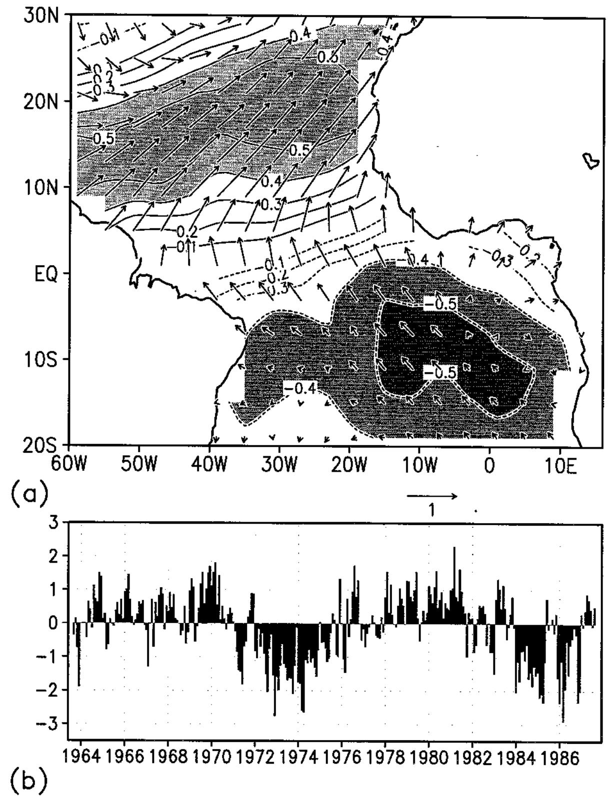


Figure 6. The dominant pattern of tropical Atlantic variability revealed by a joint EOF analysis of surface wind stress and SST monthly anomalies, September 1963–August 1987. Spatial structures of the first EOF mode and the associated time coefficients are illustrated in (a) and (b), respectively. Contours represent SST loading values, interval of 0.1°C. Vectors depict wind stress loadings. The unit vector represents an easterly wind stress anomaly of 1.0 dyn cm^{-2}. The time coefficients have been normalized by their own standard deviations. From Nobre and Shukla (1996), reproduced by permission of the American Meteorological Society

The tropical atmosphere, and especially its surface cross-equatorial flow, is sensitive to changes in interhemispheric SST gradients (see Chiang *et al.*, 2001). The NAO may also be influenced by subtropical SST anomalies. Déqué and Servain (1989) discuss teleconnections between the tropical Atlantic and SST and mid-latitude geopotential height anomalies. Sutton *et al.* (2000) show that there is a significant model response to the tripole SST pattern of Figure 2(a), and argue that it is the subtropical SST anomalies that are the most important, although the response to component parts of the tripole do not appear to be additive. It has even been suggested (Rajagopalan *et al.*, 1998; Tourre *et al.*, 1999) that there is a link between the NAO and tropical SST in the South Atlantic.

2.3. Meridional overturning circulation

The meridional overturning circulation of the ocean plays a key role in poleward transport of water properties, such as heat and freshwater, carbon and nutrients. In concert with meridional atmospheric fluxes, ocean transports balance the earth's global heat and hydrologic budgets. At 25°N, the Atlantic Ocean carries some 1.2 PW of heat northwards: approximately 60% of the net poleward ocean flux and 30% of the total flux by ocean and atmosphere at this latitude (Hall and Bryden, 1982). This poleward heat flux is intimately associated with the watermass transformations that take place as thermocline waters move north and are ultimately converted by air–sea interaction into cold North Atlantic Deep Water (NADW).

The NAO, because it is the controlling factor in variability of air–sea and ice/freshwater fluxes between the Arctic and the north Atlantic—see Coachman and Aagaard (1988), Roach *et al.* (1995) and

Weingartner *et al.* (1998)—is a primary modulator of the water mass transformation process, and perhaps the strength of the MOC and hence variability in ocean heat transport. In particular it orchestrates convection in the Greenland Sea and particularly the Labrador Sea—see Lab Sea Group, 1998. The high northern latitudes and the ocean fluxes that connect them to adjacent seas are plainly not the only constituent parts of the problem. The MOC is driven globally by up-welling, down-welling and a strong component of upper-ocean wind-forcing. Fluctuations in any one of these components might affect the strength of the MOC (see, for example, Toggweiler and Samuels, 1995 for the role of the Southern Ocean wind field, or Latif *et al.*, 2000 for the role of the tropics in re-establishing the MOC under greenhouse warming conditions; see also Cane and Clement, 1999). Nonetheless buoyancy loss in the northern high latitudes, largely controlled by the NAO/AO, is still of a fundamental importance.

3. IMPACTS

The phenomena outlined above have major impacts on the populations and environment of the Atlantic sector, as summarized in Plate 1, Table I and references therein (note that not all references listed in Plate 1 and Table I are explicitly mentioned in the text). We now review these climatic consequences in more detail.

3.1. North Atlantic Oscillation

3.1.1. Temperature and global warming. The NAO exerts a dominant influence on the wintertime temperatures of the Northern Hemisphere (see Plate 1, created using the NAO index of Hurrell, 1995a,b—a similar pattern of temperature correlation is obtained using the AO index; see Thompson and Wallace, 1998). Surface air temperature and SST in wide regions across the North Atlantic basin, in eastern North America, the Arctic, Eurasia and the Mediterranean, are significantly correlated with NAO variability. Changes in temperature over land—see, e.g. Benner, 1999—(and related changes in rainfall and storminess, see below) are of serious consequence to a wide range of human activities.

Changes of more than 1°C in the DJF-averaged surface temperature are associated with a one standard deviation change in the NAO index over the northwest Atlantic and extend from northern Europe across much of Eurasia (Hurrell, 1995b; Hurrell and van Loon, 1997, see Plate 1(a) and Figure 7). Temperature changes over Northern Africa, the Middle East and the southeast US are also notable. Of particular interest is the contribution of NAO variability to the recent trend in global/hemispheric mean wintertime temperature. Hurrell (1996) demonstrated that both NAO and ENSO are linearly related to this climate trend, with NAO dominating a larger land area than ENSO. Thompson and Wallace (1998) and Wallace *et al.* (1998) also show that an index of the AO (see Section 2.1) is strongly correlated to the temperature trend over Eurasia.

3.1.2. Precipitation and storms. Changes in mean circulation patterns over the North Atlantic are accompanied by pronounced shifts in the storm tracks and associated synoptic eddy activity. These affect the transport and convergence of atmospheric moisture and can be directly tied to changes in regional wintertime precipitation (Figure 8, see also Rogers and van Loon, 1979; Bradley *et al.*, 1987; Lamb and Peppler, 1987; Hurrell, 1995b; Beniston, 1997; Appenzeller *et al.*, 1998). Drier-than-normal conditions occur during high NAO index winters over much of central and southern Europe, the northern Mediterranean countries, and west North Africa. At the same time, wetter-than-normal conditions occur from Iceland through to Scandinavia. Over North America the effect of the NAO on precipitation is not strong. However, an out of phase relationship between the NAO index and snowfall over New England was found in a recent study by Hartley and Keables (1998). Signals in eastern North American precipitation and runoff also show a relation to the NAO (Perreault *et al.*, 1999), which could impact the development of water management rules in water resource systems.

The NAO-related changes in precipitation patterns have directly affected many European economies, largely because of the long stretch of consecutive winters in which the NAO continued to intensify in the

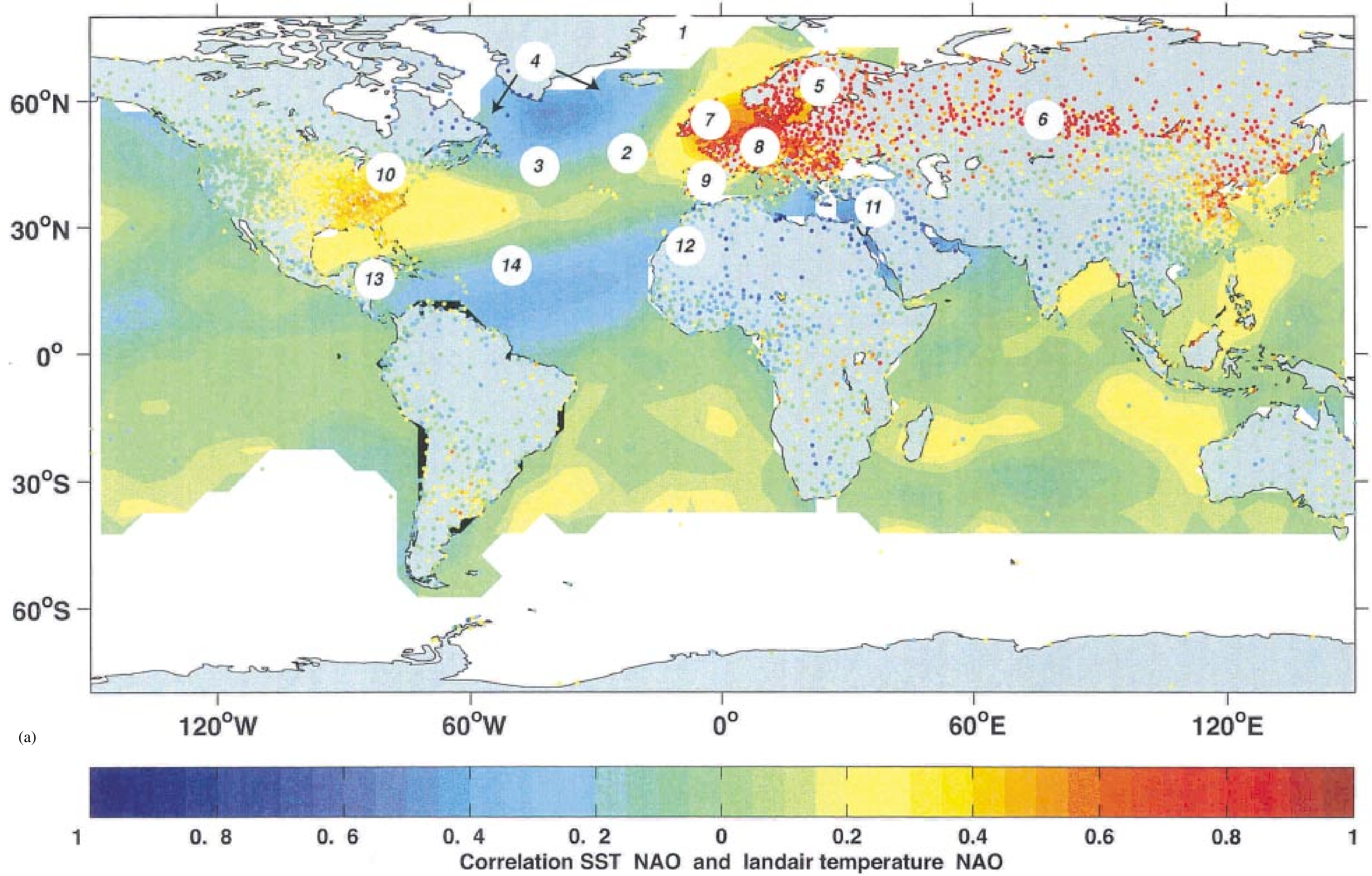
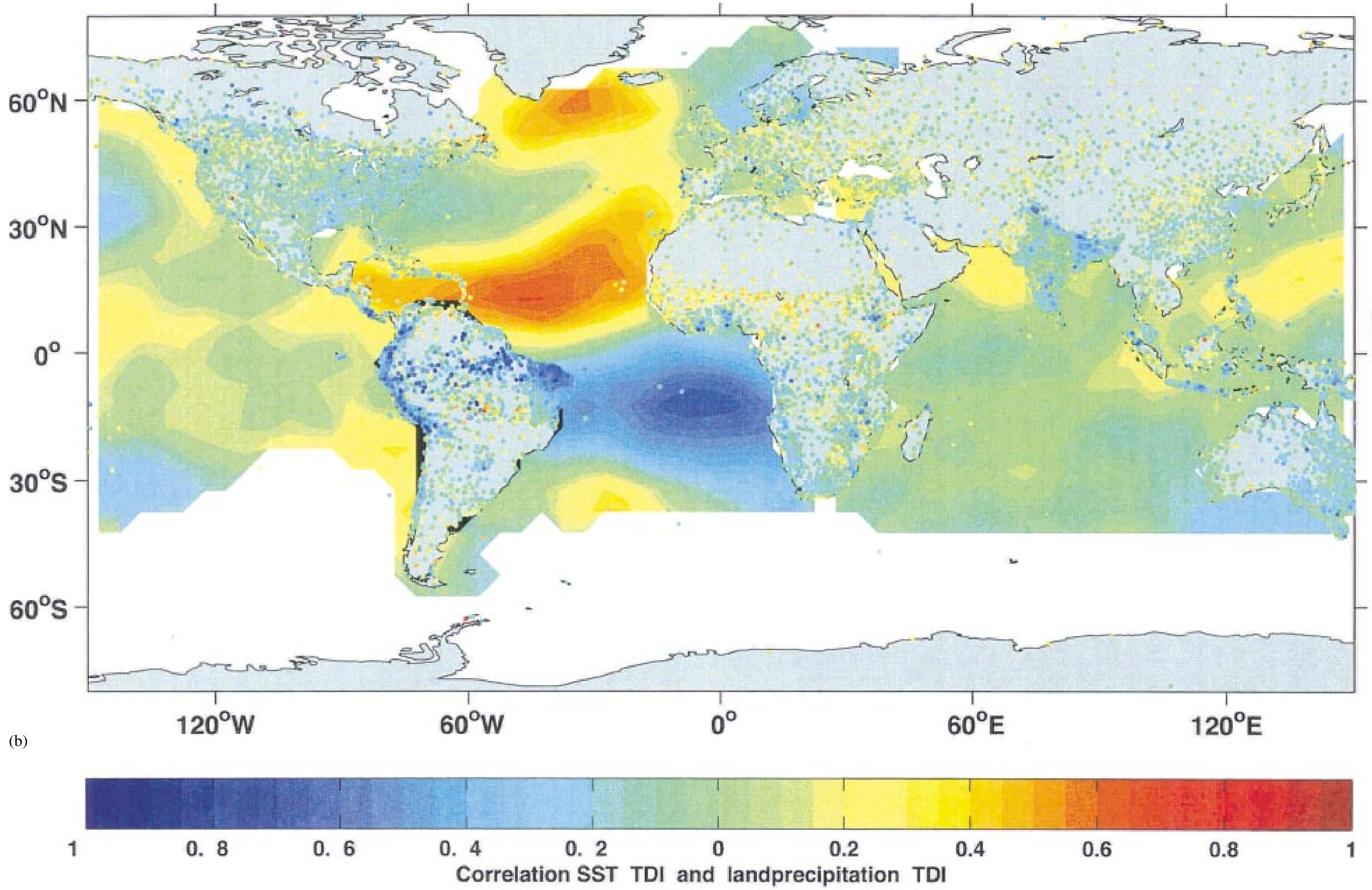


Plate 1. Impacts of Atlantic climate variability. (a) shows the correlation between the December–March NAO index (Hurrell, 1995b) and global surface temperature. SST is used over the oceans, and station air temperature over land. SST is taken from Kaplan *et al.* (1998) and station data are from NOAA/NCDA/GHCN (Vose *et al.*, 1992; Peterson and Vose, 1997). Numbers in grey circles refer to the entries in Table I (see left column). (b) shows the correlation between an index of the tropical Atlantic cross equatorial SST gradient (also known as the TDI, see Chang *et al.*, 1997) and rainfall (over land) and SST (over the oceans). SST data are taken from Kaplan *et al.* (1998) and rainfall from NOAA/NCDC/GHCN station data. All anomalies are annual (i.e. calculated from the difference between annual averages derived from monthly data and the long-term annual mean climatology)



(b)

Plate 1 (*Continued*)

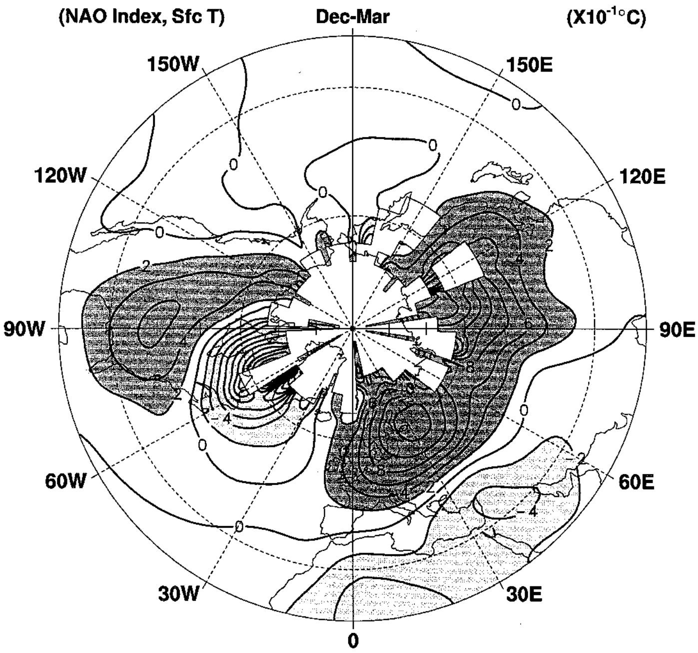


Figure 7. Illustrative changes (departures) in land surface and sea surface temperatures ($\times \frac{1}{10}$°C) corresponding to a unit deviation of the NAO index (Hurrell, 1995a,b) for the winter months December–March (1935–1998). The contour increment for surface temperature (SfcT) is 0.2°C, and regions of insufficient data are not shaded. The combined surface temperature data are those used in the IPCC assessments (also see Jones and Briffa, 1992; Parker *et al.* 1994). When the NAO is high we observe, for example, raised surface temperatures over Eurasia because the warming influence of the ocean is advected more strongly downstream. Conversely, northern Canada is anomalously cold. The reverse is true in low NAO conditions

latter part of the 20th century. Over the Alps, for example, snow depth and duration in the mid- to late 1990s were among the lowest recorded last century, causing economic hardship to those industries dependent on winter snowfall (Beniston, 1997). Severe drought also prevailed throughout Spain and Portugal affecting olive harvests. In contrast, increases in wintertime precipitation over Scandinavia led to positive mass balances in the maritime glaciers of southwest Norway, one of the few regions of the globe where glaciers have not been retreating. The effect of the NAO on precipitation has had important implications on hydroelectric power generation in, for example, Norway, normally an exporter of electricity. In 1996, a low NAO index winter (Figure 1(b)), conditions were much drier than in previous years, and concerns were voiced in Norway about meeting even national needs.

Storminess in the mid-latitudes is associated with the path of synoptic disturbances and their frontal systems. It is well known that changes in low-frequency patterns are associated with changes in storm activity and shifts in the storm tracks (Lau, 1988; Rogers, 1990). The NAO is clearly linked to systematic changes in storm tracks over the Northern Hemisphere from North America to Eurasia and the Mediterranean (Rogers, 1997).

There is a connection between the seasonally averaged NAO and low frequency variability within a season (e.g. blocking) over the Atlantic (see Nakamura, 1996). There is weaker intraseasonal variability near Greenland and the Labrador Sea and stronger intraseasonal variability over Europe, when the NAO index is positive (and vice versa).

The ocean integrates the effects of storms in the form of surface waves, to exhibit a marked response to long lasting shifts in the storm climate. The recent rising trend in the intensity of the NAO has had a profound effect on the surface wave climate of the North Atlantic. The trend in measured wave heights documented by Carter and Draper (1988) and Bacon and Carter (1991) have been supported by many anecdotal reports about increasing wave heights encountered by mariners and North Sea oil rig operators. Recent studies by Bacon and Carter (1993) and Kushnir *et al.* (1997) have associated in a more

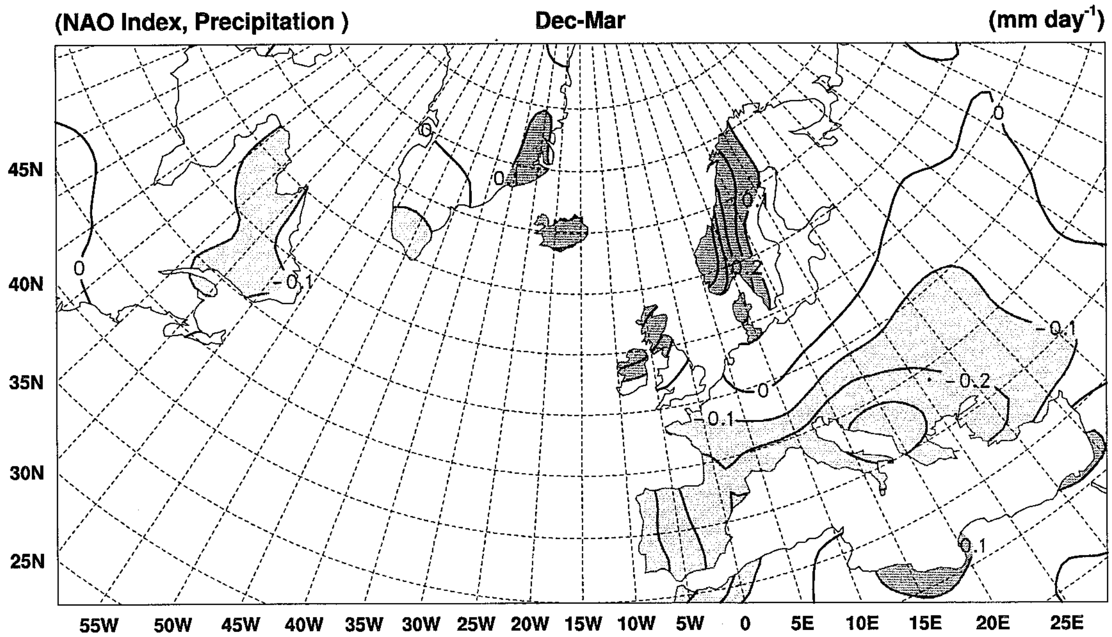


Figure 8. Illustrative changes in precipitation (mm day^{-1}) corresponding to a unit deviation of the NAO index computed over winters (December–March) from 1900 through to 1994 based on station data. When the NAO is high, storms track on a more northerly trajectory leading to anomalously high precipitation over northern Europe and anomalously low precipitation over southern Europe

quantitative manner the increase in wave heights with the increase in wintertime storminess and mean wind speeds in the North Atlantic during the last 30 years or so.

3.1.3. Fisheries and ecosystems. Changes in the NAO have also been associated with a wide range of effects on the marine ecosystem. These include changes in the production of zooplankton and the distribution of fish (e.g. Dickson and Brander, 1993; Mann and Drinkwater, 1994; Fromentin and Planque, 1996—from which Figure 9 is derived, outlying a possible connection between NAO and fish recruitment in the Labrador Sea).

NAO and fish recruitment

(+) NAO Winter Index (–)

strongly positive	strongly negative
cold air, strong winds	warm air, weak winds
large ice cover	small ice cover
large cold intermediate layer	small cold intermediate layer
low surface and bottom temperature	higher surface and bottom temperature
low surface summer salinity	high surface summer salinity
poor recruitment and growth	good recruitment and growth

Figure 9. A hypothesis concerning consequences of strongly positive or strongly negative NAO indexes on the recruitment and growth of northern Cod on the Labrador Shelf; see Mann and Drinkwater (1994)

Evidence has also recently emerged that the NAO influences temporal and spatial variability in timing of plant growth in Scandinavia: the length of the plant growth season varied by 20 days between extremes of the NAO index (Post *et al*., 1997). Through effects on vegetation and climatic conditions, the NAO was furthermore observed to influence several aspects of life history and ecology of terrestrial, large mammalian herbivores, including phenotypic variation, fecundity, demographic trends and population dynamical processes (Post *et al*., 1997). Influences of the NAO are evident among five species of ungulates in populations in Greenland, Canada, the US, the UK, Norway and Finland.

3.1.4. Sea ice. Variability of Arctic sea ice has been studied by many investigators (e.g. Walsh and Johnson, 1979; Mysak and Power, 1991; Chapman and Walsh, 1993; Fang and Wallace, 1994). A good synthesis of the interannual Arctic sea ice variability and its link to atmospheric forcing is given in a recent article by Deser *et al*. (2000). They show that in the cold season, Arctic sea ice variability occurs in the North Atlantic sector and is strongly linked to the NAO (see Figure 10). These sea ice fluctuations

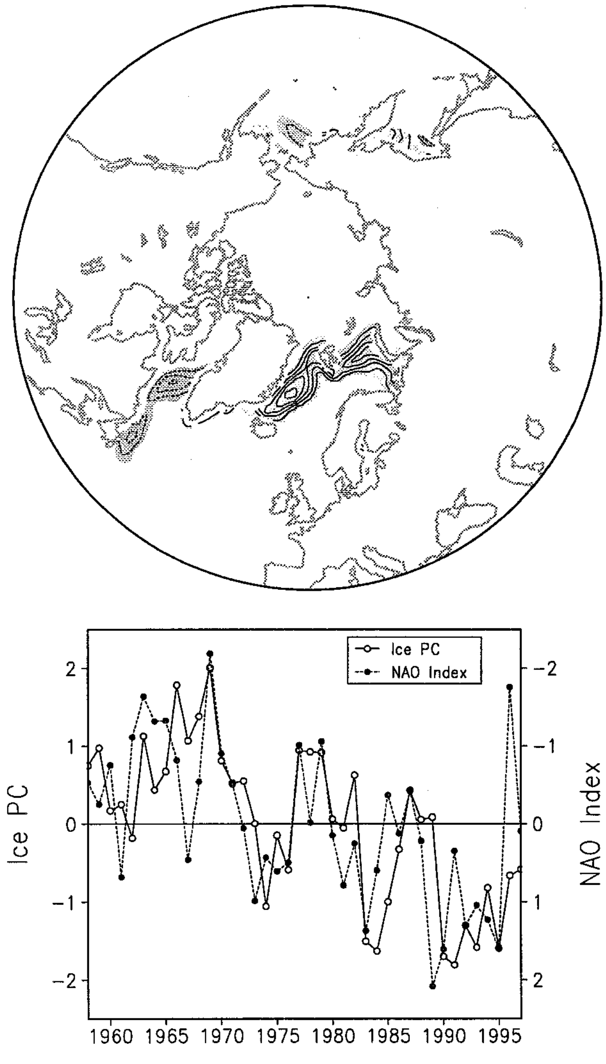


Figure 10. Top panel: leading EOF of winter sea ice concentration anomalies during 1958–1997. Contour interval is 5%, zero contour has been omitted and negative anomalies are dashed. Bottom panel: standardized time series of the first EOF of sea ice shown above (solid line) and the wintertime NAO index (scale inverted). Thus, when the NAO is low, winter sea ice concentration is anomalously high over Greenland and anomalously low over the Labrador Sea. Both panels are from Deser *et al.* (2000), reproduced by permission of the American Meteorological Society

display a seesaw pattern in ice extent between the Labrador and Greenland Seas: when wintertime SLP is lower than normal over the North Atlantic, the Labrador Sea ice boundary tends to extend further south in spring (the season of maximum sea ice extent) and the Greenland Sea ice boundary stays north of its climatological extent.

3.2. Tropical Atlantic variability

3.2.1. Rainfall and droughts. Although sea surface temperature anomalies in the tropical Atlantic are weaker than those associated with the Pacific El Niño, they are thought to play a role in sometimes disastrous climate hazards over the Americas and Africa. In the tropical Atlantic, rainfall is governed by the annual migration of the subtropical high pressure cells on both sides of the equator and the changes in their strength, as well as the north–south swings of the ITCZ. Many of the continental regions bordering the tropical Atlantic experience a sharp seasonal contrast in rainfall in which half the year is wet and the other half is dry. Some have abundant rainfall, while others are semiarid. It is in these semiarid regions that a delay in seasonal rainfall or a significant drop in its total amount can bring serious hardship; for example, subSaharan Africa (the Sahel, 15°W–15°E, 15–20°N) with its boreal summer rainfall, and the northeast corner of Brazil (Nordeste, 35°–45°W, 2°–10°S), where it rains in the boreal winter. There has been much study of the dynamics of climate anomalies in these two regions. Hastenrath and collaborators (Hastenrath and Heller, 1977; Hastenrath, 1978; Moura and Shukla, 1981; Hastenrath *et al.*, 1984; Ward and Folland, 1991; Hastenrath and Greischar, 1993; Nobre and Shukla, 1996) found that the tropical Atlantic SST distribution and associated anomalies in SLP and wind are leading factors in determining the anomalies in seasonal Nordeste rainfall. These same climatic factors were also found to be strongly linked with rainfall anomalies over the Sahel and western tropical Africa (Lamb, 1978; Hastenrath *et al.*, 1984; Folland *et al.*, 1986; 1991) and over the Caribbean/Central American sector (Enfield and Elfaro, 1999; Giannini *et al.*, 2000).

Rainfall anomalies in the Nordeste display a wide range of time scales with clear decadal components (Figure 11(top)). Such long time scales are clearly present in the Sahel rainfall time series (Figure 11(bottom)). The correlation between seasonal rainfall in the Nordeste and tropical Atlantic rainfall indicates an out of phase relationship between SST anomalies in the ocean regions underlying the Northern and Southern Hemisphere trade wind region, respectively (Hastenrath and Heller, 1977; Moura and Shukla, 1981). Dry Nordeste years tend to occur when SSTs north of the equator are higher than normal and SSTs south of the equator are less than normal (see also Wagner, 1996; Carton, 1997). Rainfall in subtropical West Africa is more complex but also displays considerable dependence on a similar out of phase variation of SST in the tropical Atlantic (Lamb, 1978; Lough, 1986; Lamb and Peppler, 1987). It rains more when Northern Hemisphere subtropical SST anomalies are positive, and Southern Hemisphere SST anomalies are negative. Recent attempts to understand the dynamics of this asymmetric SST distribution are described in Carton *et al.* (1996), Nobre and Shukla (1996) and Chang *et al.* (1997)—see Section 4.2.

3.2.2. Hurricanes. Interannual variability of the seasonal frequency of Atlantic hurricanes is related to, amongst other factors, the sign and amplitude of the SST anomaly in the subtropics. Moreover, Gray (1990) suggests that tropical cyclone intensity is correlated with the multi-decadal fluctuation of North Atlantic SST (cf. Kushnir, 1994). Gray, making inferences using empirical methods, anticipates a return to a stormier tropical climate, similar to that of the 1950s, when the North Atlantic finally recovers from its recent cold phase. Such an increase would threaten the densely populated coastal regions in the Gulf of Mexico and the North American Atlantic seaboard, which have developed markedly during the relatively benign 1970s and 1980s.

3.3. MOC and abrupt climate change

Air–sea interaction over the Atlantic—both fluxes and sea ice extent—is orchestrated by the NAO/AO, and so secular shifts between high index and low index states play a role in and can be affected

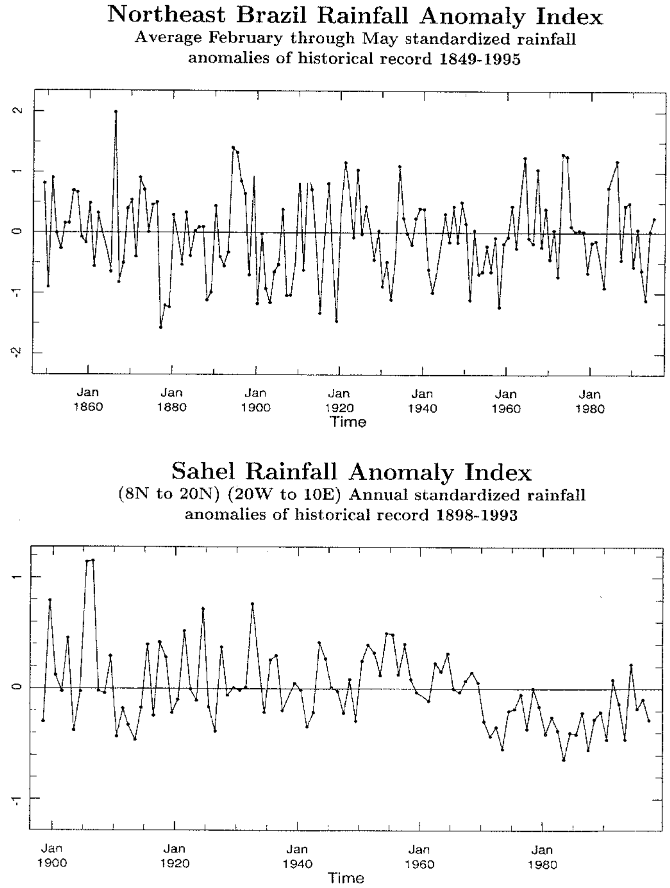


Figure 11. Northeast Brazil rainfall index (top panel) is calculated from data for Fortaleza (3.7°S, 38.5°W) and Quixeramobim (5.3°S, 39.3°W) Brazil as obtained from the NCAR World Monthly Surface Station Climatology. The Sahel series is based on an average of annual standardized anomalies from 14 stations within the region 8°–20°N and 20°W–10°E obtained from the NCAR World Monthly Surface Station Climatology

by abrupt transitions of the MOC. The role of the Atlantic in a world of increasing CO_2 emission has been the focus of much recent attention. Models suggest that under greenhouse warming, the atmosphere may move to a state in which the NAO/AO is anomalously high. Most (but not all—see Gent, 2001) projections of greenhouse gas-induced climate change anticipate a weakening of the MOC in the North Atlantic in response to increased freshening and warming in the subpolar seas—see Rahmstorf (1999), Rahmstorf and Ganopolski (1999) and Delworth and Dixon (2000). Since the overflow and descent of cold, dense waters across the Greenland–Scotland Ridge is a principal means by which the deep ocean is ventilated and renewed, the suggestion is that a reduction in upper-ocean density at high northern latitudes will weaken the MOC. If this were to occur, northern Europe and the northeastern American continent would rapidly cool.

Palaeoclimate records show that massive and abrupt climate change has occurred in the Northern Hemisphere, especially during and just after the last Ice Age—see Broecker and Denton (1989), Broecker (1997, 2000) and Marotzke (2000)—with MOC change as a plausible driver. Both paleoclimate records and models suggest that the changes in the strength of the MOC may occur rapidly, in a few decades.

In the admittedly short modern records of ocean variability, there is growing evidence that hydrographic changes of decadal scale in the Arctic and subarctic seas are able to feed south across the deep northern overflows to cause hydrographic changes in the deep and abyssal layers of the Labrador Sea (Dickson RR, Yashayaev I, Meincke J, Turrell B, Dye S, Holfort J. 2001. Rapid freshening of the deep North Atlantic over the past four decades. *Nature*, submitted). These variations are large and long sustained.

Unfortunately, our models do not yet adequately deal with many of the mechanisms believed to control the MOC, and our observations cannot yet supply many of the numbers they need. One can thus justifiably question the accuracy of model projections. However, because the paleo record suggests that the strength of the MOC has been very different in past climates, and the effects would be so dramatic, a collapse of the MOC must be taken seriously even if the chance of it happening were to be small.

4. EMERGING HYPOTHESES

Climate variability over the North Atlantic involves interaction between the troposphere, the stratosphere, the underlying ocean, adjacent land masses, the arctic to the north, the tropics to the south and remote forcing from the Pacific. Elements of the jigsaw puzzle are presented schematically in Figure 12. The Atlantic basin is rather small and comes under the influence of all these processes. Despite this complexity, it is clear that the NAO is the key and primary source of variability for North Atlantic climate on many time scales. It reaches from the tropics to the arctic, extends high up in the stratosphere and through its effect on air–sea fluxes, is a major source of variability in SST, ocean gyres and MOC. Variability in the off-equatorial tropical Atlantic is, to a significant degree, energized by the NAO (together with remote forcing from the Pacific ENSO). The confined east–west extent of the tropical Atlantic is not conducive to strong ENSO-like coupled modes that dominate variability in the Pacific, but coupled interactions may exist.

There is not yet a consensus about the relative contribution of proposed mechanisms on Atlantic climate variability—whether it be interaction with the stratosphere, tropical influences or interaction

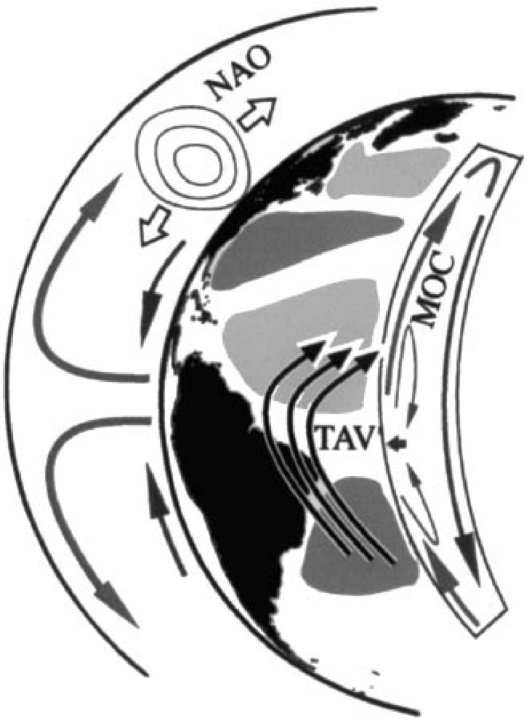


Figure 12. A schematic of TAV $\leftrightarrows$ NAO $\rightleftarrows$ MOC interactions. The NAO, reaching down into the tropics and up to high latitudes, is an important source of variability for TAV and the MOC. In turn, the TAV through its influence on tropical/subtropical SSTs can feedback on the NAO through the Hadley circulation. The MOC, a major contributor to meridional heat transport, can also affect the magnitude of the pole–equator temperature gradient over the Atlantic sector, the strength of the mid-latitude jet stream and hence the NAO. Extending high up into the stratosphere, the NAO may also influence, and be influenced by, the strength and position of the polar stratospheric vortex. The strength of the coupling between the NAO and the stratosphere above and the ocean below is not yet clear

with the ocean below. Here, therefore, we review mechanisms that have been proposed, and go on to a critical evaluation of them, hinting at what we think might be happening. This will lead us, in Section 5, to an assessment of our ability, or otherwise, to anticipate and predict the impacts reviewed in Section 3.

4.1. The North Atlantic Oscillation

4.1.1. Intrinsic atmospheric processes.

Storm tracks. The primary forcing mechanism of the NAO is thought to be eddy–mean flow interaction on synoptic time scales, i.e. dynamics internal to the atmosphere. The weather systems not only drive large-scale flow but are organized by it (Branstator, 1995). Eddy–momentum fluxes readily project on to equivalent barotropic structures, such as the NAO (Hoskins, 1983; Hoskins *et al.*, 1983; Illari, 1983; Branstator, 1992, 1995) because they too peak at tropopause level. The basic mechanism may be inextricably linked to zonally averaged dynamics—see the zonal flow vacillation studies of James and James (1992) and comments by Feldstein and Robinson (1994). Atmospheric general circulation models forced with climatological SSTs—see, e.g. Saravanan (1998)—readily display NAO-like fluctuations with a white spectrum of variability. If the fundamental NAO mechanism arises from processes internal to the atmosphere, how can we account for the redness of the observed NAO?[1] Three possible mechanisms are downward influences from the stratosphere, forcing from the tropical Atlantic and interaction with the underlying ocean.

Stratosphere/troposphere interaction. Atmospheric processes alone might produce strong interannual and longer-term variations in the intensity of the NAO through its connection to the stratosphere. Variations in the strength of the winds in the upper troposphere and lower stratosphere can condition the ‘waveguide’ affecting the upward propagation of planetary waves from the winter troposphere and, moreover, perhaps the level in the atmosphere at which those waves break. Westerly momentum, pumped into mid-latitudes by the breaking waves, could then ‘bring’ upper level winds down towards the surface. In this way an increasingly strong polar stratospheric vortex could actively excite the tropospheric NAO. If this mechanism is operative, then the increasing NAO of recent decades could be driven by (perhaps anthropogenically forced) changes in the lower stratosphere.

Whatever the mechanism, a strong statistical connection exists between the strength of the stratospheric cyclonic winter vortex and the tropospheric circulation over the North Atlantic—see Perlwitz and Graf (1995), Kodera *et al.* (1999). The strengthening of the stratospheric vortex in recent decades may have been due to, for example, tropical volcanic eruptions, ozone depletion or global warming (Perlwitz and Graf, 1995; Thompson and Wallace, 1998). In particular, Perlwitz and Graf (1995) hypothesize that stratospheric increases in CO_2 result in enhanced radiative cooling of the polar stratosphere in winter, leading to a strengthening of the polar vortex. They then invoke stratosphere–troposphere interaction to explain the recent positive trend in the NAO index through enhancement of tropospheric stationary waves. This hypothesis was recently supported by a GCM integration forced with increased CO_2 concentration (Shindell *et al.*, 1999). Baldwin and Dunkerton (1999) present observations which hint at stratospheric circulation anomalies propagating downwards to the Earth’s surface, where they are reflected as changes in the magnitude and sign of the NAO/AO.

Tropical forcing. The NAO could be remotely forced from the tropics. The Hadley Circulation is known to be sensitive to meridional SST gradients in the tropics and could act as an atmospheric ‘bridge’ to mid-latitudes, thus modulating mid-latitude NAO variability. It is not yet clear whether the Hadley cell is being orchestrated globally or whether independent regional mechanisms are coming into phase with one another. The tripole of SST associated with the NAO—see Figure 2(b)—has a strong signal in the tropics–subtropics and so variability of the jet stream could be excited from its southern flank—see, for example, Hoskins and Sardeshmukh (1987), Robertson and Mechoso (1999), Venzke *et al.* (1999). Direct forcing of the subtropical Atlantic by ENSO also occurs—see Section 4.2.

[1] One could argue that the NAO spectrum is indistinguishable from white noise (see the confidence level plotted in Figure 3(a)) and so there is nothing, perhaps, to explain. See the Discussion in Wunsch (1999) and Stephenson *et al.* (2000).

4.1.2. Interaction with the ocean.

Ocean mixed layer. An elegant model of how a mode of atmospheric variability like the NAO and the ocean mixed layer might interact was provided by Barsugli and Battisti (1998), building on the model of Frankignoul and Hasselmann (1977). They argue that on seasonal time scales, when the ocean mixed layer is in thermal equilibrium with atmospheric forcing, the heat exchange at the atmosphere–ocean interface is reduced compared with the hypothetical case when SST is not allowed to vary in response to thermal forcing. This may act to reduce thermal damping of the NAO, and according to Barsugli and Battisti is the primary effect of ocean–atmosphere coupling in the mid-latitudes. In this way thermal coupling with the ocean mixed layer enhances the persistence and variance of modes of variability such as the NAO.

It is difficult to find direct evidence of this mechanism in the observations; one cannot deduce the temporal characteristics of the NAO from observations in the absence of SST anomalies, and thus we do not have a reference. Nevertheless, there is evidence that a significant fraction of the winter NAO variance (about 25%) may be predicted from the preceding large scale SST pattern, suggesting an oceanic influence (Czaja and Frankignoul, 1999; in press). The SST pattern responsible for this forcing projects significantly onto the SST tripole forced by the NAO, thus providing a mutual positive feedback. As noted in Section 2.1.3, the SST tripole has a significant winter to winter persistence, possibly due to the re-emergence mechanism. This, combined with the positive feedback of SST on the NAO, could explain the observed 1-year memory of the winter NAO.

There is a weak and often inconsistent response in atmospheric models to mid-latitude SST anomalies (Palmer and Sun, 1985; Kushnir and Held, 1996; Peng *et al*., 1997; Venzke *et al*., 1999), which is perhaps indicative of an ambiguous response of the atmosphere too. The mechanisms at work are discussed in Peng and Whitaker (1999) and Robinson (2000). Nevertheless, Rodwell *et al*. (1999) are able to excite an NAO-like response in the Hadley Centre model with an SST tripole forcing of the form, as shown in Figure 2. Moreover, they show that when observed (global) SSTs are used as a lower boundary condition, then the low-frequency component of modelled and observed NAOs track one another over a 50-year period, if ensembles are combined. These results could be interpreted to suggest that if the ocean plays a role in setting SST, then its inherent predictability (see Section 5) might endow low-frequency aspects of the NAO with predictability too. Similar results, obtained with a different model, are reported by Mehta *et al*. (2000).

Although the Rodwell *et al*. results are of great interest, they do not necessarily imply that the NAO has predictability or that the ocean is behaving in anything other than a passive manner. Bretherton and Battisti (2000) show, using Barsugli and Battisti's model, that the behaviour observed in the Rodwell *et al*. and Mehta *et al*. integrations can be understood in terms of a weak response of the mid-latitude atmosphere to underlying SST without invoking a role for ocean circulation. These calculations and their interpretation for atmospheric predictability are also discussed in Czaja and Marshall (2000) who, however, do not discount a possible role for ocean air circulation.

Ocean circulation. We expect an imprint of the NAO on the ocean and covarying patterns of climate signals in the two fluids, because

(I) Changes in the NAO are reflected in marked changes in surface stress, air–sea heat fluxes, freshwater fluxes and STT—see Section 2.1.3.

(II) The ocean acts as an integrator of the high-frequency forcing acting on it. If an SST anomaly is 'tied' to a deep thermal anomaly, which is re-exposed each winter (Alexander and Deser, 1995; Alexander and Penland, 1996), then the SST anomaly can re-emerge year after year—a process that is known as re-emergence—and so might have a greater ability to alter, through anomalous heat fluxes, the overlying atmosphere.

(III) The interior ocean has a selective memory for winter conditions—just when the NAO is strongest—both in the subduction of winter conditions into the thermocline (Stommel, 1979; Marshall *et al*., 1993; Williams *et al*., 1995), and in the seasonal sequestering of winter conditions from below the seasonal thermocline (Qiu and Huang, 1995).

Saravanan and McWilliams (1998) generalize the Hasselmann/Frankignoul model by taking into account mean flow advection in the ocean, recognizing that although intrinsic atmospheric variability exhibits temporal incoherence, it has strong spatial coherence, which preferentially excites selected frequencies of oceanic variability through a spatial resonance. Frankignoul *et al.* (1997) invokes internal Rossby wave dynamics to select a time scale. Thermohaline circulation may also play a role. For example, multi-decadal signals in the coupled model of Delworth *et al.* (1993) appear to be associated with variations in the intensity of the thermohaline circulation and variability in the region of the Greenland Sea. Griffies and Tziperman (1995) attribute these modelled decadal fluctuations in the MOC to stochastic atmospheric forcing.

Marshall *et al.* (2001) present an idealized framework in which to consider the response of the Atlantic Ocean to, and its possible feedback on, NAO forcing—see Figure 13. They argue that a key aspect of the interaction is the meridional shift in the jet associated with the NAO which drives, through anomalous wind curl, an 'intergyre gyre', which modulates the trajectory of the Gulf Stream and North Atlantic Current. Modelling results show a signature of ocean circulation on SST on the western margin of the basin in the vicinity of the separated Gulf Stream—see, e.g. Halliwell (1998). Observations of the relation

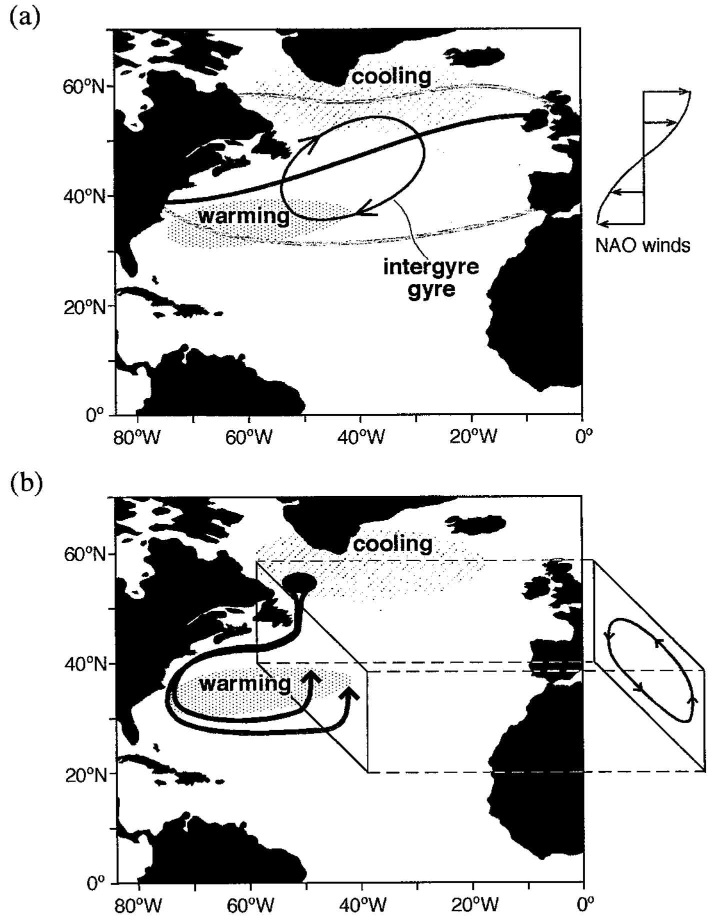


Figure 13. (a) Schematic diagram of the wind stress curl and air–sea flux anomaly patterns associated with a positive NAO. We see a 'Z' whose diagonal is the zero wind-curl line of the climatology and whose top and bottom are the zero wind-curl lines of the NAO anomaly shown in Figure 2(c). Regions of warming and cooling of the ocean due to the air–sea flux anomalies shown in Figure 2(b) are indicated. The sense of the wind-driven 'intergyre' gyre spun up by NAO(+) wind-curl forcing is also shown. (b) Schematic diagram of the anomaly in thermohaline circulation induced by the dipole in ocean thermal anomalies created by anomalies in air–sea heat fluxes associated with NAO(+) shown in Figure 2(b). The imagined anomaly in overturning circulation sketched in the meridional section on the right represents a zonal average picture

between the latitude of Gulf Stream separation and the NAO are more equivocal: Taylor and Stephens (1998) and Frankignoul *et al.* (in press) find that observed shifts do indeed lag the NAO but only by ~ 2 years.

Buoyancy effects must also play a role: when the NAO is strong, cooling of the polar oceans is enhanced; see Figure 2(b). Should this cooling persist then one might expect the vigour of the MOC to increase. Figure 13(b) also plots a schematic of the imagined anomaly in thermohaline circulation induced by the NAO(+) buoyancy fluxes. Herbaut *et al.* (2001) and Eden and Willebrand (2001) study the passive response of the Atlantic Ocean to NAO-like wind and buoyancy forcing and discuss the temporal evolution of gyres and MOC in GCMs.

4.1.3. Active atmosphere–ocean coupling. To the extent that variations in gyres (and associated instabilities) and MOC can induce variations in ocean heat transport and thence SST, the reddened spectrum of SST could then imprint itself back on the atmosphere, reddening its spectrum too. On very long (millennial) time scales, the Atlantic Ocean is indeed an active player in the climate, being responsible for a significant poleward energy transport. On intermediate (decadal?) time scales, there may be a mutual coupling between the two fluids. Atlantic decadal variability may thus, in part, reflect a coupled interaction between ocean and atmosphere in the Atlantic, in which the low-frequency response of the ocean to atmospheric forcing and its feedback on the atmospheric circulation result in low-frequency oscillations. Ocean gyres and thermohaline circulation can both play a role.

Ocean gyres. Bjerknes (1964) discussed the mutual interaction of atmosphere and ocean in the Atlantic sector in terms of the interplay of meridional heat transport in the two fluids—as sketched schematically in Figure 14 and described in the legend. He invoked anomalies in heat transport by ocean gyres and 'compensation' by atmospheric heat transport, assuming that at very long time scales the sum of atmospheric and oceanic heat transport remains constant. Bjerknes imagined the gyre anomaly to be a consequence of the state of the 'zonal index' of Rossby (1939) (equivalent to the NAO/AO) blowing over it some time in the past. This delay, which sets the time scale of the oscillation of the coupled system—it takes some 10 years for the first baroclinic mode oceanic Rossby wave to cross the Atlantic at 40°N—is

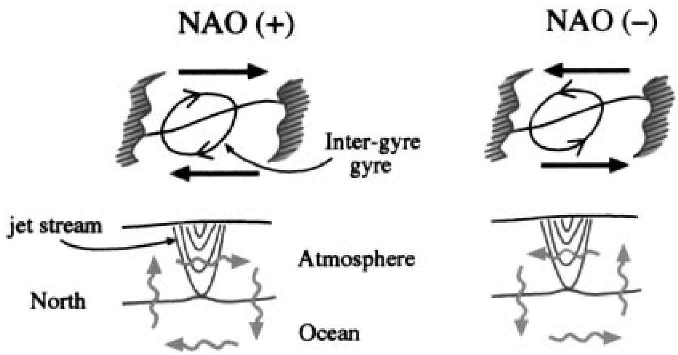


Figure 14. Schematic of anomalies in atmosphere/ocean heat transport and associated air–sea heat flux (the horizontal and vertical curly arrows) that accompany atmosphere–ocean coupling, following Bjerknes (1964). Suppose that for a period of time the NAO occupied its high index state more frequently than its low index state (solid arrows represent surface winds). Ignoring for the moment changes in ocean circulation, this would create a new equilibrium with enhanced meridional SST gradient, more heat loss from the ocean to the north and from the atmosphere (to space). However, invoking Sverdrup, Bjerknes argued that, with some delay, the North Atlantic current would have a more pronounced poleward path (the intergyre gyre begins to circulate anticyclonically), with enhanced poleward heat transport. Thus, the atmospheric jet stream will experience anomalous warming to the north and cooling to the south—the curly arrows represent this anomalous flux. The jet stream will thus weaken, it is supposed, and the anomaly in atmospheric heat transport will be to the south, compensating the enhanced northward oceanic heat transport. With weaker winds, the NAO will occupy its low index state more frequently and so, over time, the sense of the intergyre gyre will change sign to become negative, carrying heat southwards and warm the atmosphere to the south. Enhanced atmospheric temperature gradients will result in stronger winds, anomalous poleward atmospheric heat transport and an increased probability of the NAO occupying its high index state. Hence, the cycle, whose period is set by the response time of oceanic heat transport to changes in the atmospheric forcing, can repeat again

a crucial ingredient of theoretical models that couple ocean gyres with the atmospheric jet stream—see Latif and Barnett (1994), Jin (1997), Weng and Neelin (1998), Cessi (2000), Goodman and Marshall (1999), Neelin and Weng (1999), Marshall *et al.* (2001).

There is an emerging body of theory and simulation that addresses intrinsic variability of the wind-driven ocean circulation and its effect on the position of the Gulf Stream axis and SST anomalies—see, for example, Chang *et al.* (2001), Dewar (2001) and references therein. These studies suggest that ocean gyres have internal oscillatory modes of variability that can affect SST and thence, through air–sea interaction, perhaps the overlying atmosphere.

It is not known at present whether a coupled interaction between ocean gyres and the atmosphere is present in the Atlantic—there are hints that it is active in some coupled ocean models, see Grotzner *et al.* (1998) and Timmerman *et al.* (1998). Czaja and Marshall (2001) present observations of enhanced power in their Gulf Stream ΔT_{GS} index and the Greenland–Iceland Low (GIL) in the 10–20-year band, and reduced power at low frequency (see Figure 3(b)). They suggest and develop simple illustrative models that are associated with a coupled interaction between GIL, a sensitive indicator of the path of Atlantic storms, and the difference in SST across the separated Gulf Stream.

Thermohaline circulation. On decadal time scales, variability could be governed by processes that modulate the strength of the meridional circulation and its associated heat transport, and hence SST and thence the overlying atmospheric circulation—see, e.g. Timmerman *et al.* (1998). For example, one might equally well suppose that changes in thermohaline circulation—see Figure 13(b)—play a role in ocean heat transport analogous to that of changes in the strength of ocean gyres depicted in Figure 14. Dipoles in thermal anomalies generated by dipole NAO heat fluxes—see Figure 2(b)—can induce anomalies in the MOC and anomalies in its heat transport, as sketched in Figure 13(b). Thus, the curly arrows in Figure 14 could also be thought to represent anomalous heat transport by thermohaline circulation induced by NAO forcing, as discussed in Marshall *et al.* (2001). The degree to which multi-decadal variability is purely oceanic in nature, however, as suggested by Delworth *et al.* (1993), or part of a coupled mode is unclear—see Delworth and Greatbatch (2000), Selten *et al.* (1999), Hakkinen (2000).

Spall (1996) suggests that the southward propagation of LSW may have an impact on the stability characteristics of the western boundary current and influence the downstream intensity of the Gulf Stream, a region that is particularly important for air–sea interaction. This offers a mechanism by which the deeper branches of the thermohaline circulation can impact, on relatively short time scales, SST and hence the atmosphere. Curry *et al.* (1998) and Joyce *et al.* (2000) review observational evidence and develop a simple conceptual model of the delay between the transport of LSW at the Gulf Stream separation point and its creation by convection in the Labrador Sea.

Delworth and Mann (2000) argue, and present evidence from coupled GCM experiments, that variations in the strength of the thermohaline circulation play a role in SST variability on multi-decadal time scales (~ 70 years). On these long time scales, the tripole pattern of SST variability typical of the Northern Hemisphere may give way to a basin-scale interhemispheric SST anomaly, as suggested by Kushnir (1994; see also Schlesinger and Ramankutyy, 1994). Folland *et al.* (1986) noted multi-decadal time scale variations in a similar pan-Atlantic SST pattern (dipolar across the equator) with connections to Sahel droughts.

4.2. Tropical Atlantic variability

Elements of proposed mechanisms responsible for tropical Atlantic variability are represented schematically in Figure 15.

4.2.1. Remote forcing.

NAO. One very important remote source of excitation is the NAO reaching down into the tropics from mid-latitudes. A possible interpretation of the covarying fluctuations of SST and trade winds shown in Figure 6(a) is that it primarily reflects the subtropical aspect of the NAO (fluctuations in the strength and location of the Azores High). Indeed, the SST tripole forced by the NAO, Figure 2(a), and the SST pattern

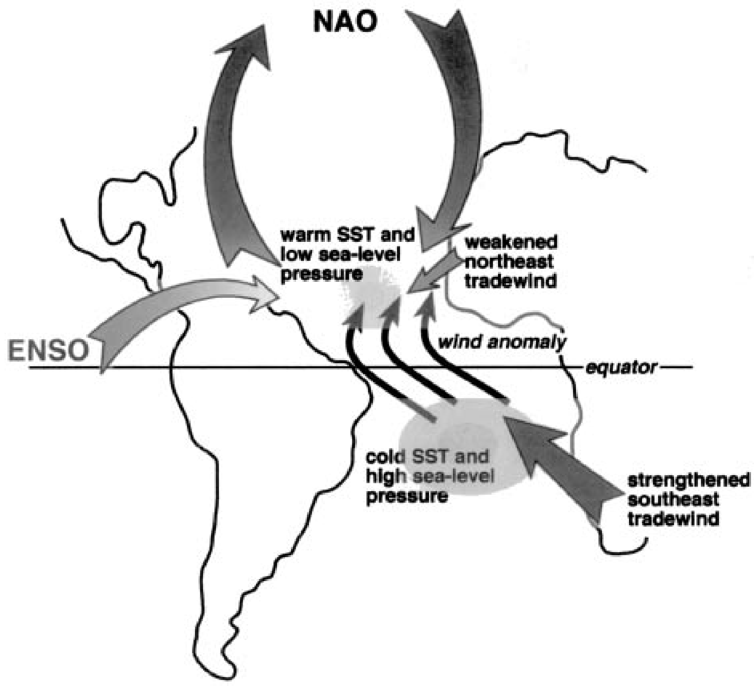


Figure 15. SST anomalies north of the equator are subject to remote influence from the NAO itself and from ENSO-related variability in the tropical Pacific. In response to a positive north–south SST difference anomaly, northward cross-equatorial winds shift the position of the ITCZ, act to reduce the northeasterly mean trade winds in the north and enhance the southeasterly trades in the south. The resulting anomalous heat flux tends to reinforce the initial north–south SST difference, which in turn strengthens the cross-equatorial wind anomalies. Negative feedbacks, perhaps due to horizontal heat transport by ocean currents, may counteract the unstable air–sea interaction yielding a self-sustained oscillation. The resulting variability of the ITCZ could influence the NAO to the north through rearranging the Hadley circulation

of Figure 6(a) are made up of common SST anomalies between the equator and 30°N. There is thus a strong coherence between the tripole and ΔT_{EQ} in the decadal band (see their power spectrum in Figure 3(b)). Xie and Tanimoto (1998) have incorporated this NAO forcing of the tropical Atlantic into a simple model, and have been able to reproduce the time evolution of the SST dipole over the past 50 years.

ENSO. The equatorial Pacific may influence the tropical Atlantic via communication along the equator, which modulates the intensity of the rising branch of the Hadley cell over Central and South America. The response is thought to be passive on ENSO time scales. Observational studies suggest that tropical Atlantic SST and wind fields are regularly affected by Pacific ENSO on an interannual time scale (Hastenrath *et al*., 1987; Hameed *et al*., 1993; Nobre and Shukla, 1996; Enfield and Mayer, 1997). In particular, Enfield and Mayer (1997) show that the ENSO-induced Atlantic SST fluctuations lag their Pacific counterparts by 4–5 months. The origin of this lag is not clear, but Klein *et al*. (1999) argue that the El Niño warming signal is communicated to the tropical Atlantic via a reduction in surface latent heat flux associated with reduced trade winds. Recent analyses of two different GCMs (Saravanan and Chang, 1999; Sutton *et al*., 2000) have successfully isolated the ENSO-related response of the tropical Atlantic in their models.

4.2.2. Feedback processes over land. Feedback processes over land may play a role in the persistence of anomalies in rainfall in sub-Saharan Africa and other regions. Charney (1975) and Charney *et al*. (1977) propose a 'bio-geophysical feedback mechanism' in which deterioration of the vegetation cover over land can lead to a reduction of surface net radiation, which in turn enhances subsidence and reduces cloudiness and precipitation. The decreased rainfall would further diminish the vegetation cover. Courel *et al*. (1984)

present evidence of decreased albedo during the dry season of the Sahel, which is consistent with changes in plant cover. These land feedback processes need to be further examined in a fully coupled system.

4.2.3. Coupling with the ocean.
Interhemispheric SST anomalies. Sutton *et al.* (2000) and Chang *et al.* (2000a) provide clear confirmation of the effect of tropical Atlantic SST anomalies on the atmospheric circulation in two different GCMs. They show that the increase in cross-equatorial SST gradient strongly affects the cross-equatorial low-level tropospheric flow on the western side of the basin (see also the study of Chiang *et al.*, 2001). When SST is warmer than normal in the northern tropics or cold in the south, there is a marked southerly wind anomaly along the Brazilian seaboard. This is accompanied by a northward shift in the ITCZ and a reduction in rainfall during the boreal spring season. This suggests that the circulation on both sides of the equator is related. However, the fundamental driver is the variability of the interhemispheric SST gradient; the mechanism does not require that SST changes in each hemisphere to be simultaneous.

It has been hypothesized (Carton *et al.*, 1996; Chang *et al.*, 1997; Xie, 1999; Seager *et al.*, 2000; Chang *et al.*, 2000b) that decadal variations of the interhemispheric SST gradient may stem from regional ocean–atmosphere interaction involving wind–evaporation–SST (WES) positive feedback: a reduction of evaporation caused by a decrease in total wind speed. Carton *et al.* (1996) demonstrate that the wind-induced latent heat flux acts to enhance SST variability both north and south of the equator in the tropical Atlantic Ocean. Chang *et al.* (1997) further suggest that there is a mutual interaction between the wind-induced heat flux and SST. In response to a positive north–south anomaly in SST difference, northward cross-equatorial winds act to reduce the northeasterly mean trade winds in the north and enhance the southeasterly trades in the south. Invoking WES, they argue that anomalous heat flux tends to reinforce the initial north–south SST difference, which in turn strengthens the cross-equatorial wind anomalies. Negative feedbacks, perhaps due to horizontal heat transport by ocean currents and/or atmospheric radiative and turbulent energy fluxes, counteract the unstable air–sea interaction, yielding a self-sustained oscillation. Decadal time scales can be obtained but the period is highly sensitive to model parameters.

Subtropical-tropical interactions. Gu and Philander (1997) hypothesize that exchange between extratropical and tropical water masses through thermocline ventilation can induce decadal changes in the depth of the equatorial thermocline in the Pacific, which in turn can modify equatorial wave speeds and give rise to decadal SST variability. Hansen and Bezdek (1996) presented observational evidence of anomalous SST variability on decadal scales extending into the tropical Atlantic along the preferred path of the subtropical gyre. Malanotte-Rizzoli *et al.* (2000) discuss and model water–mass pathways between the subtropics and the tropics in the Atlantic. Shallow exchange windows from the northern subtropics ($\sim 30°$N) are found and the important role of the seasonal cycle and its effect on the position of the ITCZ highlighted. It has even been suggested—see Yang (1999)—that on a decadal scale, TAV might be linked to convection in the Labrador Sea. However, the full extent to which these tropical–extratropical ocean exchange processes affect Atlantic SST variability is unknown.

The Atlantic ENSO. The Atlantic ENSO mode appears to be stable (Zebiak, 1993) as anticipated from delayed oscillator theory (Battisti and Hirst, 1989). Hence, the Atlantic ENSO is likely to be forced by external stochastic forcing. Delecluse *et al.* (1994) suggest that the Pacific ENSO provides one possible source of external forcing for interannual SST variability in the eastern equatorial Atlantic. NAO forcing from the subtropics is another.

4.3. Critical evaluation of mechanisms

It is clear that the NAO is a mode of variability internal to the atmosphere whose primary excitation mechanism is the interaction of the Atlantic jet stream with synoptic eddies. The NAO power spectrum has increasing amounts of power at lower frequencies, a characteristic that is more noteworthy than local peaks. This redness is unlikely to be a consequence of the internal dynamics of the NAO, and is suggestive of an 'external influence'. The most likely influences are the stratosphere, the underlying ocean and the tropics.

If the NAO is best thought of as a regional manifestation of a hemispheric phenomenon—the AO—then through its influence on the vertical/meridional propagation of planetary waves, it may be sensitive to the strength of the stratospheric polar vortex. It is not yet clear how much 'downward control' is going on (the stratospheric to tropospheric mass ratio is very low). Details of the mechanism and a quantification of its effect on the troposphere are still a subject of active research. But if the linkage is sufficiently strong, the redness of the NAO spectrum could be a consequence of anthropogenically induced changes in the stratosphere.

The underlying ocean is driven by stochastic-in-time but coherent-in-space air–sea flux variability that reflects the spatial pattern of the NAO. It integrates over the temporal 'noise' of the forcing producing a 'red' power spectrum in oceanic variables. Observations suggest that SST does have a small but discernible impact on the NAO, and so could play a role in the reddening of the NAO spectrum. This small positive feedback is in agreement with modelling studies. It is difficult to isolate localized patterns of SST to which the atmosphere is most sensitive, but basin-wide SST patterns that resemble the SST tripole itself seem to be involved with tropical/subtropical centres of action probably playing the major role. In middle to high latitudes persistence associated with the re-emergence of SST with the annual cycle, can, through its feedback on the overlying atmosphere, lead to some year-to-year persistence in atmospheric variables.

On seasonal to interannual time scales, the coupled system can be rather well described by stochastic climate models appropriately modified to allow for (i) feedback between SST and air temperature, which leads to reduced thermal damping and (ii) enhanced persistence via the re-emergence process.

On longer time scales, however, the ocean—via wind and thermohaline circulation—can modulate SST variability in the vicinity of the separated Gulf Stream on the western margin of the basin, where advection by ocean currents is particularly strong. The ocean could imprint itself back on the atmosphere on longer time scales, thus reddening atmospheric variability; this might be called 'passive coupling' in that it does not involve active coupled dynamics. In this way the ocean can modulate the amplitude and phase of the NAO on decadal time scales, but it can do so without 'active dynamical coupled atmosphere–ocean modes'. We do not exclude the possible existence of active coupled modes, but it may not be necessary to invoke them.

It seems likely that the MOC of the ocean can support damped modes of internal variability that can be excited by stochastic atmospheric forcing whose primary source is the NAO. Abrupt shifts of the MOC occur in models (and as suggested by the paleo record, perhaps in nature too), which could plausibly lead to shifts in the secular trends of the NAO and thence TAV.

In the tropical Atlantic, there does not appear to be a single dominant mechanism controlling variability. The atmosphere in the tropical Atlantic is responsive to changes in SST, and particularly the cross-equatorial gradient. A primary source of variability appears to be external forcing from the NAO and ENSO, whose effect can be enhanced through, for example, the WES mechanism. What is unclear is if the interaction involves interhemispheric, or single-hemisphere SST gradients, and to what extent ocean dynamics play a role. Along the equator local interactions seem to be important, but its predictability and influence outside the region remains undetermined.

5. DISCUSSION AND PROSPECTS FOR PREDICTABILITY

5.1. Predictability

Climate predictability in the North Atlantic region can be divided into two types:

1. Atmospheric predictability: prediction of the statistical properties of meteorological variables that affect human life directly, such as the surface air temperature and precipitation over land.

2. Oceanic predictability: prediction of the state of the North Atlantic Ocean, including surface properties such as the SST, and also the subsurface temperature, salinity, and current structure.

Although the two types of predictability are closely related, it is useful to distinguish between them. They involve very different processes and time scales and have different uses. Atmospheric predictability can only be addressed in a statistical sense, since the deterministic predictability limit is of the order of a week. Since oceanic time scales are much longer, one could conceivably make deterministic predictions for the ocean on decadal time scales.

The social utility of atmospheric predictions is clear. Oceanic predictability, as discussed below, is often a prerequisite for long-term atmospheric predictability. However, it can also be of intrinsic value in itself. The most direct application is, of course, to fisheries and marine ecosystems. There are also more indirect applications. Much effort has been put into simulating the mean state of the ocean using oceanic or coupled GCMs. However, one of the primary uses of GCMs is to simulate climate variability. Predicting the distributions of dynamical oceanic variables and biogeochemical tracers on interannual to decadal time scales using GCMs, and then testing these predictions using observations, can be a very good way to test the GCMs. Certainly tests of this sort would be essential before one can put more trust into predictions of, for example, 'thermohaline collapse' from coarse resolution ocean GCMs.

Since atmospheric processes have very short memory, to carry out atmospheric prediction on longer than seasonal time scales, one needs to invoke some kind of memory in the boundary conditions. This memory most likely resides in the ocean in the form of SST anomalies, and therefore oceanic predictability is often a necessary condition for atmospheric predictability (unless the memory happens to reside in some other component of the climate system, such as sea ice or the land surface). Observational studies of Atlantic variability have shown there are indeed long-lived SST anomalies with coherent spatial structure in the mid-latitudes (Deser and Blackmon, 1993; Kushnir, 1994; Czaja and Marshall, 2001) and in the tropics (Carton *et al*., 1996).

Another necessary condition for atmospheric predictability is that there be a coherent response in the atmospheric flow to SST anomalies. Most atmospheric GCM studies find a rather weak response to mid-latitude SST anomalies, when compared with internal atmospheric variability (Kushnir and Held, 1996; Blade, 1997; Lau, 1997; Saravanan, 1998). Although this could be a deficiency of the current generation of GCMs, it could also turn out to be a property of the real climate system. A coherent atmospheric response to SST anomalies in the tropical Atlantic could also provide atmospheric predictability. Although the local effects of tropical Atlantic SST anomalies are fairly well established, their remote effects have only begun to be addressed. Observations, however, suggest that there is some predictability in the early winter NAO—see Czaja and Frankignoul (1999). As discussed in Czaja and Frankignoul (in press), the signal is mainly from mid-latitudes but there is a competing forcing from the tropics.

It is often argued that the weak and often inconsistent response of current atmospheric GCMs to mid-latitude SST anomalies is perhaps the biggest stumbling block to progress in using numerical models to study atmospheric predictability. It should be noted, however, that the relevance to the coupled problem of prescribing a fixed SST (implying an infinite heat capacity) and seeking a response is open to question—see Sutton and Mathieu (in press). Ideally, to quantify the atmospheric predictability, one needs to address the coupled problem with higher resolution models using better physical parameterizations. While our GCMs are being improved, statistical and empirical models of atmospheric predictability could be employed.

Some AGCMs—see, e.g. Rodwell *et al*. (1999) and Mehta *et al*. (2000)—show modest skill in reproducing aspects of the observed NAO behaviour, especially its interdecadal fluctuations when forced with the time history of observed, global SSTs and sea ice concentrations over the past 50 years or so. However, as discussed previously, these integrations can be understood in terms of a weak response of the mid-latitude atmosphere to underlying SST without invoking a role for ocean circulation.

Oceanic predictability, unlike atmospheric predictability, does not require memory to reside in the boundary conditions, since oceanic processes can easily supply the memory. One can have predictability in the deep ocean even if there is no atmospheric predictability at all. Indeed, modelling work by Griffies and Bryan (1997) suggests that the North Atlantic Ocean may have predictability on the order of a decade or longer. All that is required is the presence of spatially and temporally coherent oceanic variability on interannual and longer time scales. This variability could be generated through several different mechanisms. A plausible scenario is that there are damped modes of oceanic variability that are excited by stochastic atmospheric forcing (Griffies and Tziperman, 1995; Saravanan and McWilliams, 1997; Delworth and Greatbatch, 2000; Marshall *et al*., 2001).

5.2. *Outlook*

Scientists in Europe and the US are planning a major research program focusing on Atlantic climate variability in the context of CLIVAR, an international climate research program. The major elements of this program are designed to investigate simultaneously the ocean–atmosphere–land–ice interactions associated with three major climate phenomena reviewed here: the NAO, TAV and MOC.

It is proposed to pursue a balanced approach between observations, modelling and theory as well as diagnostic studies, to understand better the primary climate phenomena, their interactions and the potential for predictability. This will be achieved by implementing a network of sustained observations, which will serve as the backbone for diagnostic studies and coupled data assimilation. Initial observational activities will cover the Atlantic with a focus on the sector north of 30°S, and include the gateways to the Arctic Ocean.

A series of shorter term regionally focused experiments are planned, which will be imbedded within the sustained observing network to allow in-depth understanding of particular regions and/or key processes. Historical and proxy datasets will be enhanced and oceanographic and atmospheric reanalyses conducted. It is hoped that the program will lead to (i) an Atlantic sector component of a global climate observing system that will be able to constrain coupled climate models; (ii) improved coupled climate models that are able to simulate the observed climate variability of the last few decades and, hence, will be the basis for future climate predictions on seasonal to multi-decadal time scales; and (iii) collect comprehensive datasets of observed and/or analysed fields that form the basis for Atlantic climate variability studies.

The main objectives of the program are to:

- Describe and model coupled atmosphere–ocean–land interactions in the Atlantic sector, quantify their influences on the regional and global climate system, and determine their predictability.
- Assemble quantitative historical, proxy and real time datasets that may be used to test, improve and initialize models of coupled Atlantic climate variability.
- Investigate the sensitivity of the MOC to changes in surface forcing and assess the likelihood of abrupt climate change.

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