CHAPTER 2

Climate variability over the North Atlantic

James W. Hurrell and Robert R. Dickson

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2.1 Introduction

The climate of the Atlantic sector and surrounding continents exhibits considerable variability over a wide range of timescales. It is manifest as coherent fluctuations in ocean and land temperature, rainfall, and surface pressure with a myriad of impacts on society and the environment. Of central importance is the North Atlantic Oscillation (NAO), which dictates much of the climate variability from the eastern seaboard of the United States to Siberia and from the Arctic to the subtropical Atlantic, especially during boreal winter. The NAO refers to a redistribution of atmospheric mass between the Arctic and the subtropical Atlantic, and swings from one phase to another to produce large changes in the mean wind speed and direction over the Atlantic, the heat and moisture transport between the Atlantic and the neighbouring continents, and the intensity and number of storms, their paths, and their associated weather. Such variations have a significant impact on the wind- and buoyancy-driven ocean circulation, as well the site and intensity of water mass transformation, so that the strength and character of the Atlantic meridional overturning circulation (MOC) is substantially influenced.

In this chapter, we provide a broad review of the NAO and its forcing of the NAO. Of particular interest is the long, irregular amplification of the oscillation towards one extreme phase during winter over recent decades. This climatic event, which is unprecedented in the modern instrumental record of NAO behaviour, has produced a wide range of effects on North Atlantic ecosystems. Some attention will also be given to the climatic impacts of periods of atypical NAO behaviour, such as the spatial displacement of the main centres of action in some winters, or to periods when other patterns of large-scale Atlantic climate variability are more pronounced. An in-depth treatment of the full range of Atlantic climate variability is beyond the scope of this chapter; however, the interested reader is encouraged to pursue the many references to scientific works included herein. For a comprehensive and multidisciplinary overview of material (theory, observations, and models) related specifically to the NAO, the reader should also consult *The North Atlantic Oscillation: Climatic Significance and Environmental Impact* (2003).

2.2 The North Atlantic Atmosphere

2.2.1 Mean state

The long-term (1899–1999) distribution of sea-level pressure (SLP) over the Northern Hemisphere (NH) is illustrated in Fig. 2.1. Large changes from boreal winter (December–February) to boreal summer (June–August) are evident. Perhaps most noticeable are those over the Asian continent related to the development of the Siberian anticyclone during winter and the monsoon cyclone over southeast Asia during summer. Over the northern oceans, subtropical anticyclones dominate during summer, with the Azores high-pressure system covering nearly all of the North Atlantic. These anticyclones weaken and move towards the equator by winter, when the high-latitude Aleutian and Icelandic low-pressure centres predominate.

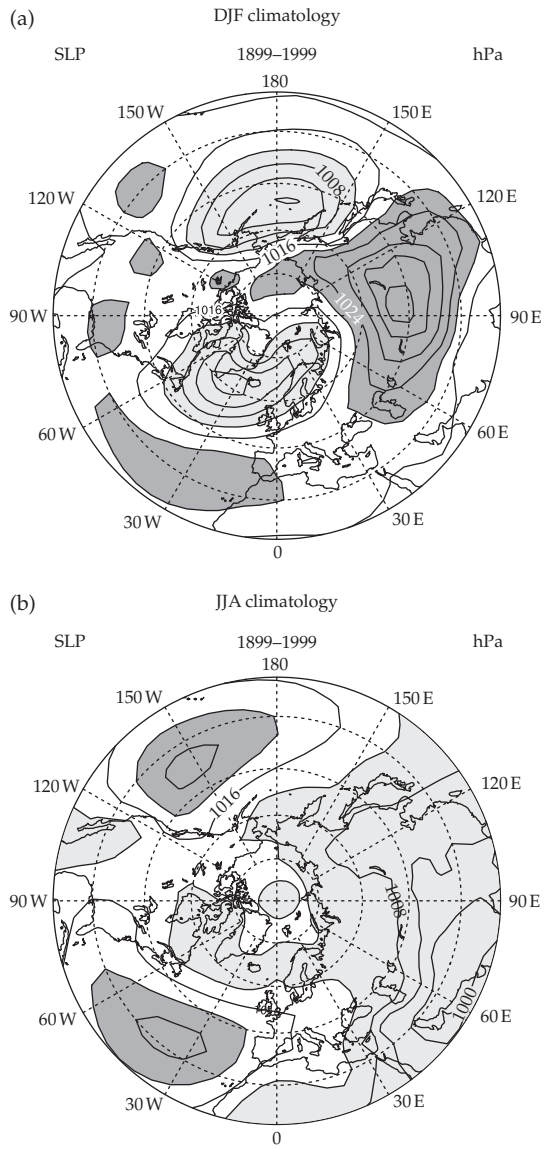

Figure 2.1 Mean (1899–1999) distribution of SLP (hPa) over the NH during boreal winter (December–February) and summer (June–August). The contour increment is 4 hPa, and values less (greater) than 1012 hPa (1024 hPa) are indicated by the light (dark) shading.

Because air flows counterclockwise around low pressure and clockwise around high pressure in the NH, westerly flow across the middle latitudes of the Atlantic occurs throughout the year. The vigour of the flow is related to the pressure gradient, so the surface winds are strongest during winter when they average between 5 and 10 m s^{-1} from the eastern United States across the Atlantic onto northern Europe. The middle latitude westerlies extend throughout the troposphere and reach their maximum (up to 40 m s^{-1}) at a height of about 12 km. This 'jet stream' roughly depicts the path of storms (atmospheric disturbances operating on timescales of about a week or less) travelling between North America and Europe. Over the subtropical Atlantic the prevailing northeasterly winds are relatively steady but strongest during boreal summer.

2.2.2 Variability

What is the NAO?

Monthly mean surface pressures vary markedly about the long-term mean SLP distribution (Fig. 2.1). This variability occurs in well-defined spatial patterns (Wallace and Gutzler 1981; Barnston and Livezey 1987) particularly during boreal winter over the NH. Such variations are commonly referred to as 'teleconnections' in the meteorological literature, since they result in simultaneous variations in weather and climate over widely separated points on earth. One of the most prominent patterns is the NAO. Meteorologists have noted its pronounced influence on the climate of the Atlantic basin for more than two centuries (van Loon and Rogers 1978).

The NAO refers to a north–south oscillation in atmospheric mass between the Icelandic low- and the Azores high-pressure centres (Walker and Bliss 1932). It is most clearly identified when time-averaged data (monthly or seasonal) are examined, since time averaging reduces the 'noise' of small-scale and transient meteorological phenomena not related to large-scale climate variability. The spatial signature and temporal variability of the NAO are usually defined through the regional SLP field, for which some of the longest instrumental records exist. It is also readily apparent in meteorological data to the lower stratosphere, however, where the seesaw in mass between the polar cap and the middle latitudes is much more zonally symmetric. This more annular mode of variability has been termed the Arctic Oscillation (AO) by Thompson and Wallace (1998). That the NAO and AO reflect essentially the same mode of surface variability is emphasized by the similarity of their time series, with differences depending mostly on the details of the analysis procedure (Deser 2000; Wallace 2000).

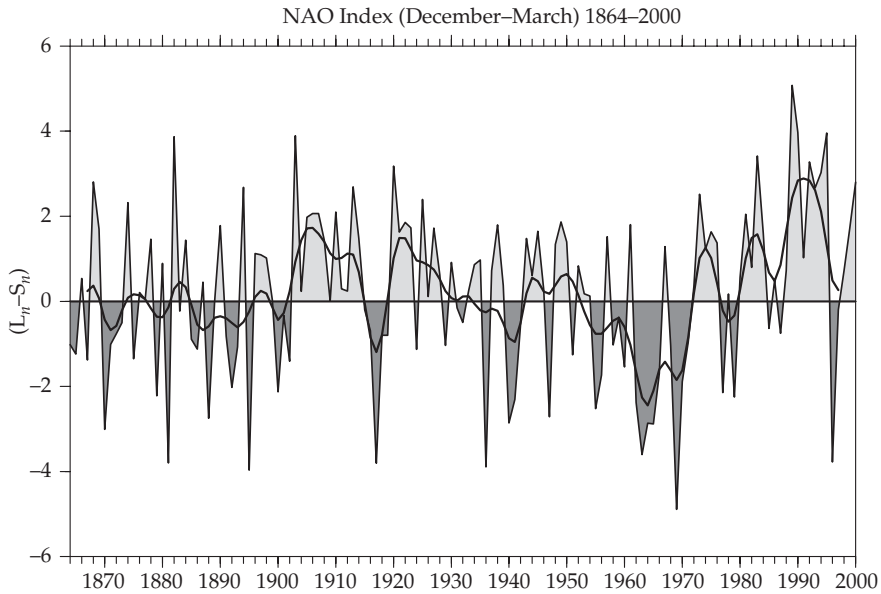

Figure 2.2 Winter (December–March) index of the NAO based on the difference of normalized SLP between Lisbon, Portugal, and Stykkisholmur/Reykjavik, Iceland from 1864 through 2000. The indicated year corresponds to January (e.g. 1950 is December 1949–March 1950). The average winter SLP data at each station were normalized by division of each seasonal pressure by the long-term mean (1864–1983) standard deviation. The heavy solid line represents the index smoothed to remove fluctuations with periods less than 4 years.

A time series (or index) of more than 100 years of NAO variability is shown in Fig. 2.2,[1] and a schematic of the spatial signature and climate impacts of the oscillation is shown in Fig. 2.3. Although the NAO is evident throughout the year, these plots illustrate conditions during boreal winter when the atmosphere is dynamically the most active. During the months of December through March, for instance, the NAO accounts for more than one-third of the total variance in SLP over the North Atlantic, substantially more than any other pattern of variability (Barnston and Livezey 1987; Rogers 1990).

Differences of more than 15 hPa in SLP occur across the North Atlantic between the two phases of the NAO in winter (Hurrell 1995a). In the so-called positive phase, higher than normal surface pressures south of 55°N combine with a broad region of anomalously low pressure throughout the Arctic and subarctic (Fig. 2.3(a)). Consequently, this phase of the oscillation is associated with stronger-than-average westerly winds across the middle latitudes of the Atlantic onto Europe, with anomalous southerly flow over the eastern United States and anomalous northerly flow across western Greenland, the Canadian Arctic, and the Mediterranean. The easterly trade winds over the subtropical North Atlantic are also enhanced during the positive phase of the oscillation. During the negative phase, both the Icelandic low- and Azores high-pressure centres are weaker-than-normal, so both the middle latitude westerlies and the subtropical trade winds are also weak (Fig. 2.3(b)).

There is little evidence for the NAO to vary on any preferred timescale (Fig. 2.2). Large changes can occur from one winter to the next, and there is also a considerable amount of variability within a given winter season (Nakamura 1996). This is consistent with the notion that much of the atmospheric circulation variability in the form of the NAO arises from processes internal to the atmosphere, in which various scales of motion interact with one another to produce random (and thus

[1] More sophisticated and objective statistical techniques, such as eigenvector analysis, yield time series and spatial patterns of average winter SLP variability very similar to those shown in Figs 2.2 and 2.3.

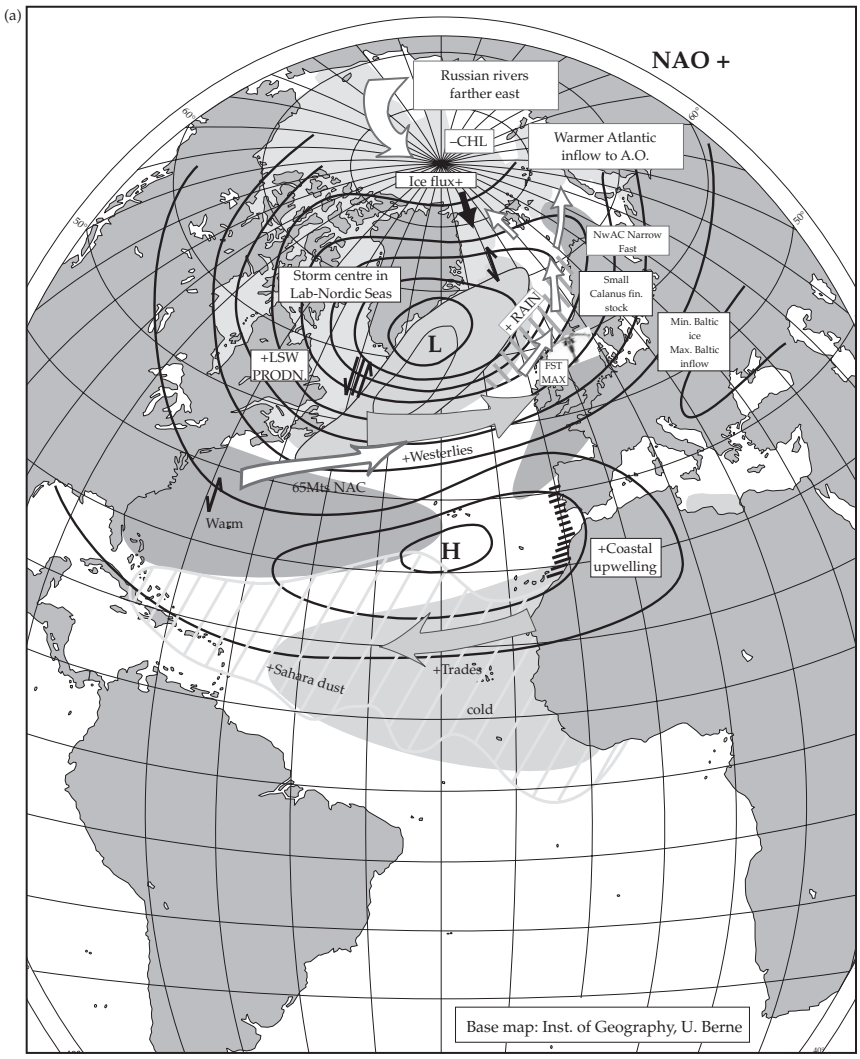

Figure 2.3 A schematic of the Atlantic–Arctic sector under (a) NAO-positive and (b) NAO-negative conditions. The following will explain the abbreviated text and symbols and provide references to the sources of information. (See Plate 1(a) and (b)).

Note:

1. Bold (weak) up-down arrows (↑ and ↕) indicate intense open (capped) ocean convection and strong (weak) heat flux. Boxes labelled '+18°W Prodn', '+LSW Prodn' and '+GSDW Prodn' indicate the location and NAO state associated with the maximal production of the three main Atlantic Mode Waters—18° Water, Labrador Sea Water and Greenland Sea Deep Water respectively. This relationship and the coordination of Mode Water formation by the NAO are described in Dickson *et al*. (1996); Dickson (1997); Talley (1996); Joyce *et al*. (2000).

2. 'Storm Centre off US Coast' and 'Storm Centre in Lab-Nordic Seas' boxes are from Rogers (1990); see also Bacon and Carter (1993); Kushnir *et al*. (1997); Alexandersson *et al*. (1998), and Carter (1999).

3. SST distributions are the cold = blue, warm = pink areas from Cayan (1992c), see also Becker and Pauly (1996), Cayan (1992a,b), Dickson and Turrell (1999).

4. 'LC + 1 SV' refers to the fact that the transport of the Labrador Current is thought to increase by 1 Sverdrup in NAO-negative extrema (Myers *et al*. 1989; see also Marsh *et al*. (1999) and Drinkwater *et al*. (1999).

5. '50 Mts^{-1} NAC' and '65 Mts^{-1} NAC' refer to the calculated increase in transport of the North Atlantic Current by 30% as NAO− changes to NAO+. The entire gyre circulation is strengthened, so that it would be equally appropriate to refer to a strengthening of the subtropical and subpolar gyre recirculations as the NAO amplifies. There is recent German evidence from their repeat hydrography during WOCE that the subpolar gyre expands eastwards as the NAO has amplified. See Curry and McCartney (2001), and also Bersch *et al*. (1999) and McCartney *et al*. (1997).

6. 'Ice flux −' and Ice flux +' in the Fram Strait refer to the observation that changes in the annual volume flux of ice from the Arctic via the western Fram Strait are associated with the NAO (Vinje and Finnekasa 1986; Hilmer *et al*. 1998; Vinje *et al*. 1998; Kwok and Rothrock 1999; Dickson *et al*. 2000). Dickson *et al*. (2000) and Hilmer *et al*. (1998) suggest this association may not have applied in earlier decades.

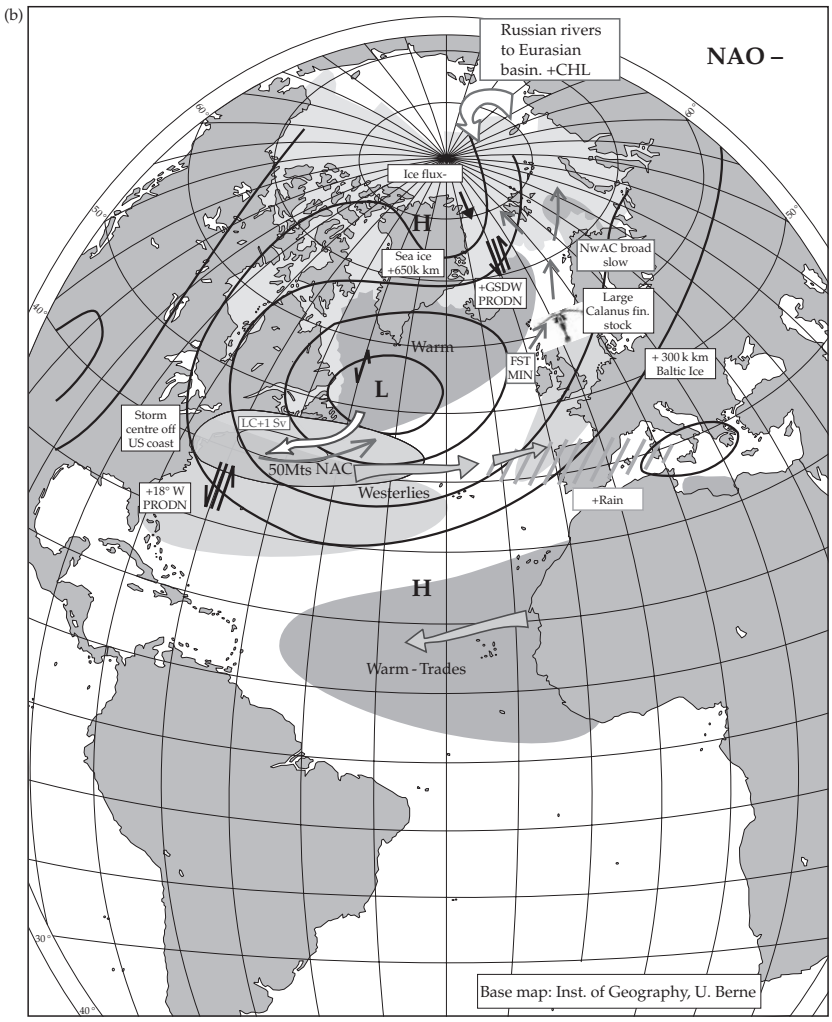

Figure 2.3 (*Continued*)

7. 'Sea ice + 650,000 km^{2}' in the 'NAO−' schematic represents the retraction of the median sea-ice border at the end of winter (April) between the most extreme 7-year runs of NAO-negative winters (1963–69) and NAO-positive winters (1989–95). The distributional change is illustrated in Dickson *et al*. (2000) while the correlation between ice concentration and the NAO index is most clearly indicated by Deser *et al*. (2000).

8. '+Rain' and its changing distribution (Dickson *et al*. 2000; see also Cayan and Reverdin 1994; Hurrell 1995a; Hurrell and van Loon 1997).

9. 'Large Calanus fin. Stock' and 'Small Calanus fin. Stock' (see Fromentin and Planque 1996).

10. 'NwAC Broad slow' and 'NwAC Narrow Fast' are from Blindheim *et al*. (2000) who use the 35 isohaline to define the width of the Norwegian Atlantic Current west of Norway.

11. 'FST MAX' and 'FST MIN' refer to the Faroe—Shetland Transport as calculated in an inverse treatment of two historic Scottish hydrographic sections (Dye 1999).

12. Baltic ice and inflow, from B. Mackenzie, personal communication.

13. '+ Saharan Dust' and the dust distribution under NAO+ conditions (Moulin *et al*. 1997).

14. Changes in middle latitude westerlies and subtropical trade winds, primarily from Cayan (1992c).

15. We have chosen to use SLP-anomaly distributions rather than sea level pressure *per se* since they better explain the sense of the anomalous airflow (based on Rogers 1990; see also Walsh *et al*. 1996; Serreze *et al*. 1997; Dickson *et al*. 2000).

16. 'Warmer Atlantic Inflow to the AO' refers to the increase in the temperature, extent and perhaps volume transport of Atlantic water inflow to the Arctic Ocean (Quadfasel *et al*. 1991; Grotefendt *et al*. 1998; Morison *et al*. 1998a; Dickson *et al*. 2000; see also Carmack *et al*. 1995; Carmack *et al*. 1997; Swift *et al*. 1997; Morison *et al*. 2001).

17. 'Russian Rivers Further East', '− CHL' on NAO + and 'Russian Rivers to Eurasian Basin; + CHL' from Steele and Boyd (1998) and Wieslaw Maslowski (personal communication).

unpredictable) variations (Wallace and Lau 1985; Lau and Nath 1991; Ting and Lau 1993; Hurrell 1995b). There are, however, periods when anomalous NAO-like circulation patterns persist over many consecutive winters. In the Icelandic region, for instance, SLP tended to be anomalously low during winter from the turn of the century until about 1930 (positive NAO index), while the 1960s were characterized by unusually high surface pressure and severe winters from Greenland across northern Europe (negative NAO index). A sharp reversal has occurred over the past 30 years, with strongly positive NAO index values since 1980 and SLP anomalies across the North Atlantic and Arctic that resemble those in Fig. 2.3(a). In fact, the magnitude of the recent upward trend is unprecedented in the observational record (Hurrell 1995a; Thompson *et al*. 2000) and, based on reconstructions using paleoclimate and model data, perhaps over the past several centuries as well (Osborn *et al*. 1999; Stockton and Glueck 1999). Whether such low frequency (interdecadal) NAO variability arises from interactions of the North Atlantic atmosphere with other, more slowly varying components of the climate system such as the ocean (Rodwell *et al*. 1999; Mehta *et al*. 2000; Hoerling *et al*. 2001), whether the recent upward trend reflects a human influence on climate (Corti *et al*. 1999; Fyfe *et al*. 1999; Osborn *et al*. 1999; Shindell *et al*. 1999; Ulbrich and Christof 1999; Gillett *et al*. 2000, 2001; Monahan *et al*. 2000), or whether the longer timescale variations in the relatively short instrumental record simply reflect finite sampling of a purely random process (Wunsch 1999), are topics of considerable current interest.

2.2.3 The NAO and surface temperature variability

The NAO exerts a dominant influence on wintertime temperatures across much of the NH. Surface air temperature and sea-surface temperature (SST) across wide regions of the North Atlantic Ocean, North America, the Arctic, Eurasia, and the Mediterranean are significantly correlated with NAO variability[2] (see also the section '*direct response to recent NAO forcing at the ocean surface*').

[2] SSTs are used to monitor surface air temperature over the oceans because intermittent sampling is a major problem and SSTs have much greater persistence.

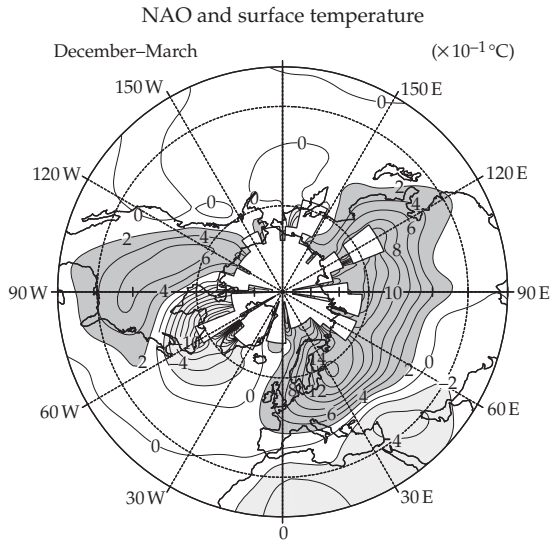

Figure 2.4 Changes in land surface and SST ($\times 10^{-1}$ °C) corresponding to a unit deviation of the NAO index for the winter months (December–March) from 1935–99. The contour increment is 0.2 °C. Temperature changes >0.2 °C are indicated by dark shading, and those <−0.2 °C are indicated by light shading. Regions of insufficient data are not contoured.

Such changes in surface temperature (and related changes in rainfall and storminess) can have significant impacts on a wide range of human activities as well as on marine and terrestrial ecosystems.

When the NAO index is positive, enhanced westerly flow across the North Atlantic during winter moves relatively warm (and moist) maritime air over much of Europe and far downstream across Asia, while stronger northerlies over Greenland and northeastern Canada carry cold air southward and decrease land temperatures and SST over the northwest Atlantic (Figs 2.4 and 2.6(b)). Temperature variations over North Africa and the Middle East (cooling), as well as North America (warming), associated with the stronger clockwise flow around the subtropical Atlantic high-pressure centre are also notable.

The pattern of temperature change associated with the NAO is important. Because the heat storage capacity of the ocean is much greater than that of land, changes in continental surface temperatures are much larger than those over the oceans, so they tend to dominate average NH (and global) temperature variability. Given the especially large and coherent NAO signal across the Eurasian

continent from the Atlantic to the Pacific (Fig. 2.4), it is not surprising that NAO variability explains about one-third of the NH interannual surface temperature variance during winter, and that the trend towards the positive NAO phase in recent decades has contributed significantly to observed global warming (Hurrell 1996; Thompson *et al*. 2000).

2.2.4 The NAO, storms and precipitation variability

Changes in the mean circulation patterns over the North Atlantic are accompanied by changes in the intensity and number of storms, their paths, and their associated weather. During winter, a well-defined storm track connects the North Pacific and North Atlantic basins, with maximum storm activity over the oceans. The details of changes in storminess differ depending on the analysis method and whether one focuses on surface or upper-air features. Generally, however, positive NAO index winters are associated with a northeastward shift in the Atlantic storm activity (Fig. 2.3(a)), with enhanced activity from southern Greenland across Iceland into northern Europe and a modest decrease in activity to the south (Rogers 1990, 1997; Hurrell and van Loon 1997; Serreze *et al*. 1997). The latter is most noticeable off the east coast of the United States and from the Azores across the Iberian Peninsula and the Mediterranean. Positive NAO winters are also typified by more intense and frequent storms in the vicinity of Iceland and the Norwegian Sea (Serreze *et al*. 1997; Deser *et al*. 2000).

Changes in the mean flow and storminess associated with swings in the NAO are also reflected in pronounced changes in the transport and convergence of atmospheric moisture and, thus, the distribution of precipitation. Anomalously low precipitation rates occur over much of Greenland and the Canadian Arctic during high NAO index winters (Fig. 2.3), as well as over much of central and southern Europe, the Mediterranean, and parts of the Middle East. In contrast, more precipitation than normal falls from Iceland through Scandinavia (Hurrell 1995a; Dai *et al*. 1997; Dickson *et al*. 2000).

This spatial pattern, together with the upward trend in the NAO index since the late 1960s (Fig. 2.2), is consistent with recent observed changes in precipitation over much of the Atlantic basin. One of the few regions of the world where glaciers have not exhibited a retreat over the past several decades is in Scandinavia (Hagen 1995; Sigurdsson and Jonsson 1995) where more than average amounts of precipitation have been typical of many winters since the early 1980s. In contrast, over the Alps, snow depth and duration in recent winters have been among the lowest recorded this century, and the retreat of Alpine glaciers has been widespread (Frank 1997). Severe drought has persisted throughout parts of Spain and Portugal as well. As far eastward as Turkey, river runoff is significantly correlated with NAO variability (Cullen and deMenocal 2000).

2.2.5 Atypical NAO winters and other atmospheric variations

While the NAO explains a substantial portion of the variability over the North Atlantic, the chaotic nature of the atmospheric circulation means that, at most times, there are significant departures from the schematic in Fig. 2.3. Even during periods of strongly positive or negative NAO index winters, the atmospheric circulation typically exhibits significant local departures from the idealized NAO pattern. For instance, Hilmer and Jung (2000) have shown that the centres of maximum interannual variability in SLP associated with the NAO have been located further to the east since the late 1970s, when the NAO winter index has mostly been positive. Such longitudinal displacements affect the NAO-related interannual variability of sea ice, surface temperature, surface heat fluxes, precipitation, and storms (see also the section '*Temporal shifts in the NAO pattern*'). Another example is the winter of 1996, which was characterized by a very negative NAO index (Fig. 2.3). Conditions during the 1996 winter over much of Europe were severe; however, the anomalous anticyclonic circulation was located well to the east of the canonical NAO pattern (Fig. 2.3(b)), with positive SLP anomalies of more than 9 hPa centred over Scandinavia. Such persistent high-pressure anomalies are a typical feature of the North Atlantic climate and are referred to as 'blocks'. The longitudes of 150°W and 15°W are particularly favoured for the development of blocking highs, which occur when the westerly flow across the middle latitudes of the NH is weak and typified by an exaggerated wave pattern. North Atlantic blocking is most typical during boreal spring and early summer, but it occurs throughout the year. Over recent decades, for instance, increased anticyclonicity and easterly flows from the warm, summer

AQ: Pls chk representation of NAO index.

continent have brought anomalously warm and dry conditions over much of northern Europe and the United Kingdom (Wright *et al*. 1999).

2.3 The North Atlantic Ocean

2.3.1 Mean state and variability

By the end of the nineteenth century, a surprisingly modern distribution of hydrographic stations had already traced out the basic physical geography of the North Atlantic Ocean and Nordic Seas. The properties of the upper ocean reflect a two-gyre structure with warm waters circulating anticyclonically in a subtropical gyre to the south and east and cold waters circulating cyclonically in a subpolar gyre to the north and west (Fig. 2.5). The main North Atlantic Current flows east along the boundary between the two gyres, where there is an abrupt change of water properties (the Oceanic Arctic Front).

Determining the *rate* of the Atlantic circulation has proved more elusive. In the absence of direct observations, evidence is largely restricted to episodes when some anomalous ocean-climate 'signal' has propagated through the system. The dates in Fig. 2.5 provide one well-known example, tracing the slow ($\approx$3 cm s^{-1}) spread of the great salinity anomaly (GSA) through the Atlantic gyre circulation in the late 1960s and 1970s (Dickson *et al*. 1988).

It is difficult, therefore, to provide separate descriptions of the mean state of the ocean and its variability since both have coloured the observational record. Here, instead, the differential response of the surface-, intermediate- and deep-layers of the North Atlantic Ocean to the recent

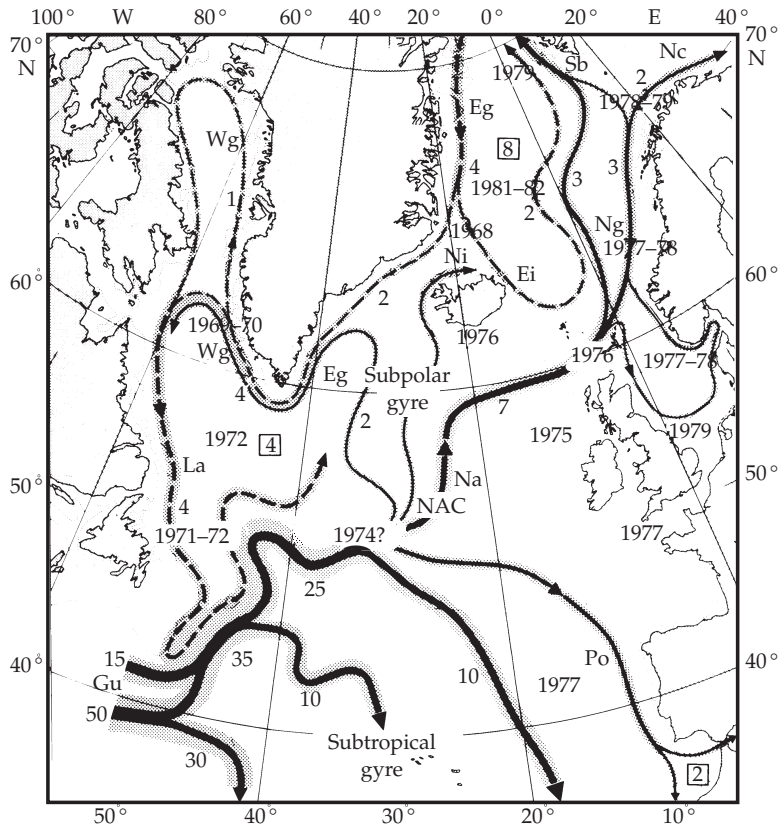

Figure 2.5 Transport scheme for the 0–1000 m layer of the northern North Atlantic with dates of arrival of the GSA superimposed (Dickson *et al*. 1988). Jakobsson (1992) concluded the GSA 'has probably generated more variability in fisheries during the last quarter of a century than any other hydrographic event in recent years'.

NAO trend (Fig. 2.2) is described, as are the responses to two episodes of atypical NAO behaviour. Most attention is focused on the northwest Atlantic and Labrador Sea, which have special climatic sensitivity and importance.

2.3.2 Direct response to recent NAO forcing at the ocean surface

It has long been recognized that fluctuations in SST and the strength of the NAO are related (Bjerknes 1964). The pattern of SST variability associated with the winter NAO consists of a tri-polar structure marked, in the positive NAO phase, by a cold anomaly in the subpolar North Atlantic, warm anomalies extending from Cape Hatteras in the west Atlantic to Biscay in the east and poleward along the Norwegian coast, and a cold subtropical anomaly between the equator and 30°N (Figs 2.3(a) and 2.6(b)). This structure suggests that the SST anomalies are primarily driven by changes in the surface wind and air–sea heat exchanges associated with NAO variations (Cayan 1992c). Indeed, the relationship is strongest when the NAO index leads an index of the SST variability by several weeks (Deser and Timlin 1997).

The pattern of correlation between the winter NAO index and scalar wind speed reflects the redistribution of storm activity described earlier (Section 2.2.4). The positive correlation between the NAO index and wind strength over much of the Atlantic midlatitudes and Nordic Seas and extending into the North Sea reflects both the northeastward extension of the storm track under NAO-positive conditions and the development of storms to maturity along that track (Figs 2.3(a) and 2.6(a)). Extreme cold and dry air streams out from the Canadian Arctic across the Labrador Sea during positive NAO index winters, increasing the monthly sea-to-air flux of sensible and latent heat by approximately 200 $W\,ms^{-1}$ compared to negative NAO index winters (Cayan 1992c). The result is the development of extremely cold SSTs in Atlantic midlatitudes from the Davis Strait and West Greenland Banks across the Labrador and Irminger Seas to Iceland and the Faroe Islands (Fig. 2.6(b)). By contrast, the warm, moist southerly airflow that is directed along the eastern boundary of the North Atlantic produces very warm SSTs in the North Sea and along the Norwegian coast to the Barents Sea, in spite of the strong wind speeds there (see also figures 7a and 8a of Cayan 1992c). West of Africa, anomalously strong northeasterly trade winds during positive NAO winters increase the surface fluxes of heat to the atmosphere, thereby cooling the ocean surface (Figs 2.3(a) and 2.6(b)).

On longer timescales, the recent upward trend towards positive NAO index winters and the corresponding northeastward extension and increase of winter storm activity (e.g. Alexandersson 1998) have been associated with an increase in wave heights over the northeast Atlantic and a decrease south of 40°N (Bacon and Carter 1993; Kushnir *et al*. 1997; Carter 1999). These changes are also illustrated by the remarkable increase in mean scalar wind speeds at Utsire in the north-central North Sea during the whole of the year (Fig. 2.7). Such changes have a range of fundamental consequences; for the vernal blooming of phytoplankton (Sverdrup 1953), for the operation and safety of shipping, for offshore industries such as oil and gas exploration, and for coastal development.

Variations in the NAO are also coupled to changes in Arctic sea ice, where the strongest interannual variability occurs in the North Atlantic sector (Deser *et al*. 2000). Sea ice fluctuations display a seesaw in ice extent between the Labrador and Greenland Seas (Fig. 2.8). Strong interannual variability is evident in the sea ice changes over the North Atlantic, as are longer-term fluctuations including a trend over the past 30 years of diminishing (increasing) ice concentration during boreal winter east (west) of Greenland (Chapman and Walsh 1993; Maslanik *et al*. 1996; Cavalieri *et al*. 1997; McPhee *et al*. 1998; Parkinson *et al*. 1998). Associated with the sea ice fluctuations are large-scale changes in SLP that closely resemble the NAO (Fig. 2.8).

When the NAO is in its positive phase, the Labrador Sea ice boundary extends farther south while the Greenland Sea ice boundary is north of its climatological mean extent. Given the implied surface wind changes (Fig. 2.3a), this is qualitatively consistent with the notion that sea ice anomalies are directly forced by the atmosphere, either dynamically via wind-driven ice drift anomalies, or thermodynamically through surface air temperature anomalies (Prinsenberg *et al*. 1997). The relationship between the NAO index (Fig. 2.2) and an index of the North Atlantic ice variations is indeed strong, although it does not hold for all individual winters (Deser *et al*. 2000). This last point illustrates the importance of the regional atmospheric circulation in forcing the extent of sea ice.

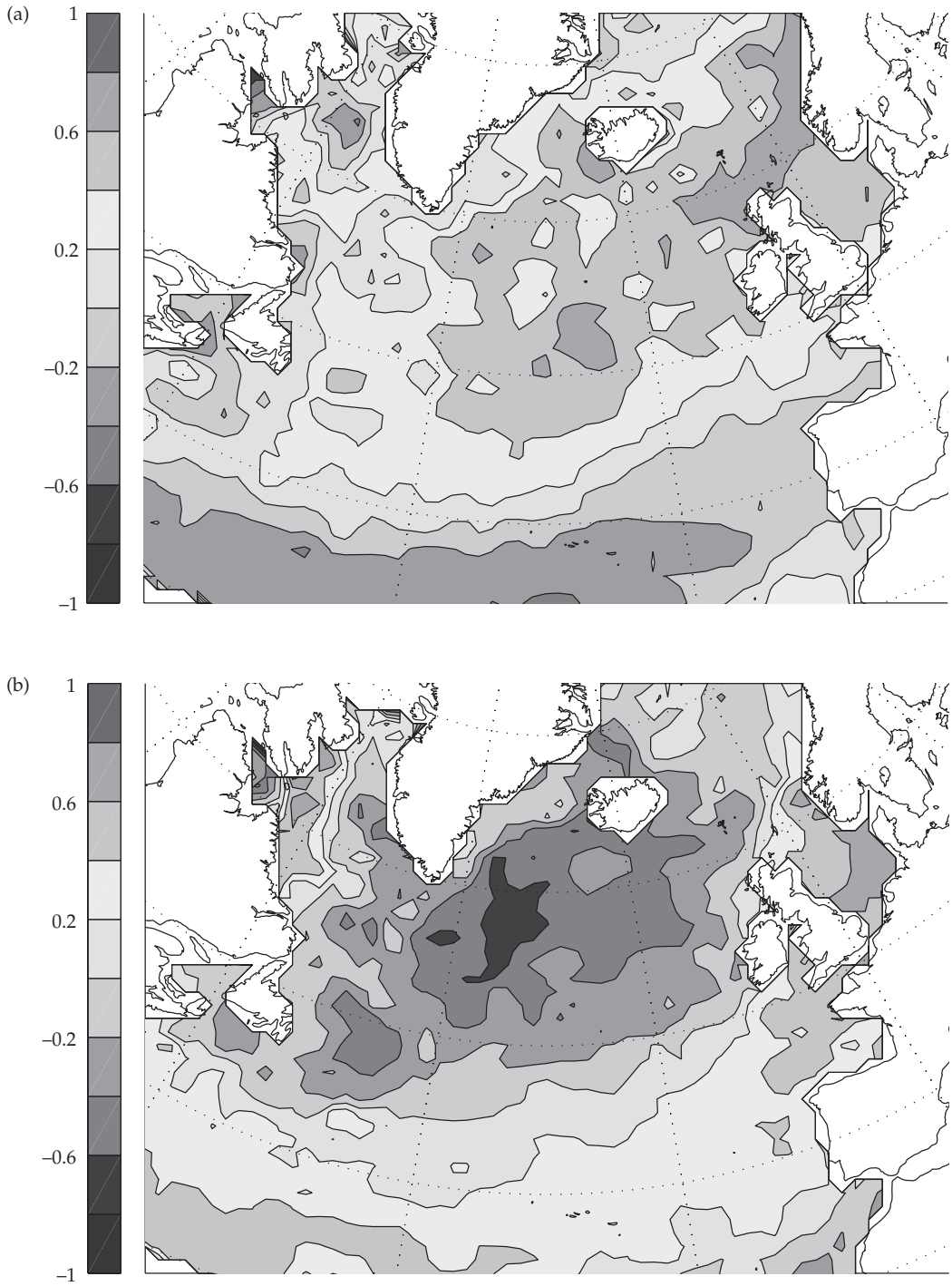

Figure 2.6 Spatial distribution of the Pearson correlation coefficient between the NAO index and (a) local scalar windspeed and (b) local SST for winter (January–March) over 1950–95 (courtesy of Ben Planque). (See Plate 2.)

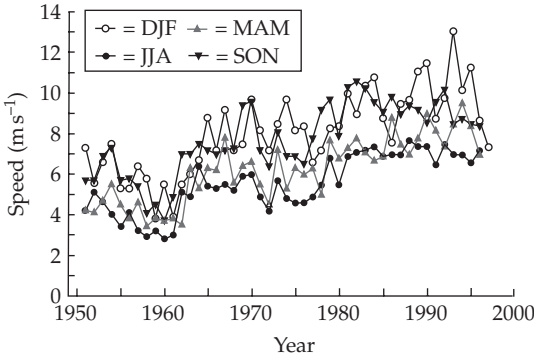

Figure 2.7 The observed increase in seasonal mean scalar windspeeds at Utsire (59 19′N 4 53′E) in the northern North Sea during the long-term amplification of the NAO between the mid-1960s and the mid-1990s.

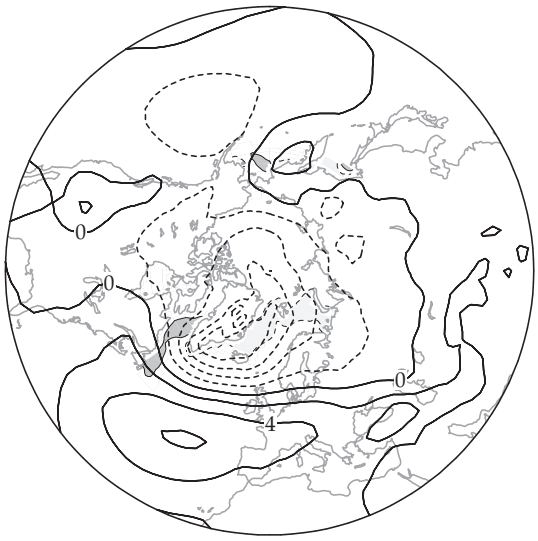

Figure 2.8 Winter SLP (contours) and ice concentration (shading) anomalies associated with the leading pattern of ice variability. The SLP anomaly field is in units of hectapascals (40 year)$^{-1}$, contoured every two units. Light (dark) shading denotes ice concentration anomalies greater than 8% (40 year)$^{-1}$ [less than −8% (40 year)$^{-1}$]. From Deser *et al*. (2000).

2.3.3 Direct and indirect response to NAO forcing at intermediate ocean depths

Subsurface ocean observations more clearly depict long-term climate variability, because the effect of the annual cycle and month-to-month variability in the atmospheric circulation decays rapidly with depth. Although these measurements are much more limited than surface observations, they reveal a decadal 'system' to the temperature evolution of the upper North Atlantic Ocean, with large organized SST anomalies circulating around the gyres coincident with decadal changes in the NAO (Hansen and Bezdek 1996; McCartney *et al*. 1997; Curry *et al*. 1998). These recurrent surface features track the movement of deep-seated anomalies formed by winter convection and the formation of vertically extensive and homogeneous intermediate-depth water masses known as 'mode waters'.

Mode waters form where the water mass structure, ocean circulation, and winter surface heat exchange favour deep-reaching convection. Since the same shifting patterns of winter storm activity and surface heat exchange that affect the surface also affect the formation of mode waters, the depth of vertical exchange also varies greatly in both time and space: to depths of 1000–3500 m in the Greenland Sea, 1000–2300 m in the Labrador Sea, and a few hundred meters in the Sargasso Sea, where 18° Water forms. The regional importance of these localized sites is that the winter renewal of mode waters provides a mechanism for carrying the signal of climate change to intermediate and greater depths, and throughout the ocean basin by horizontal spreading. It is of more than local importance also that the scale of NAO forcing appears to have imposed a degree of coordination on convective activity at the three aforementioned sites (Dickson *et al*. 1996). Deep convection over the Labrador Sea, for instance, was at its weakest and shallowest in the postwar instrumental record during the late 1960s. Since then, Labrador Seawater has become progressively colder and fresher, with intense convective activity to unprecedented depths (>2300 m) in the early to mid-1990s (Fig. 2.9). In contrast, warmer and saltier deep waters in recent years are the result of suppressed convection in the Greenland and Sargasso Seas, whereas ventilation was a maximum at these sites during the 1960s.

The reasons for the synchronization with the NAO are beginning to emerge. During the negative NAO index years of the 1960s, an enhanced land–sea temperature gradient in winter contributed to the formation of more storms than normal off the eastern seaboard of the United States (Dickson and Namias 1976; Hayden 1981). Offshore, the cold, stormy conditions caused maximum formation and ventilation of the 18° Water

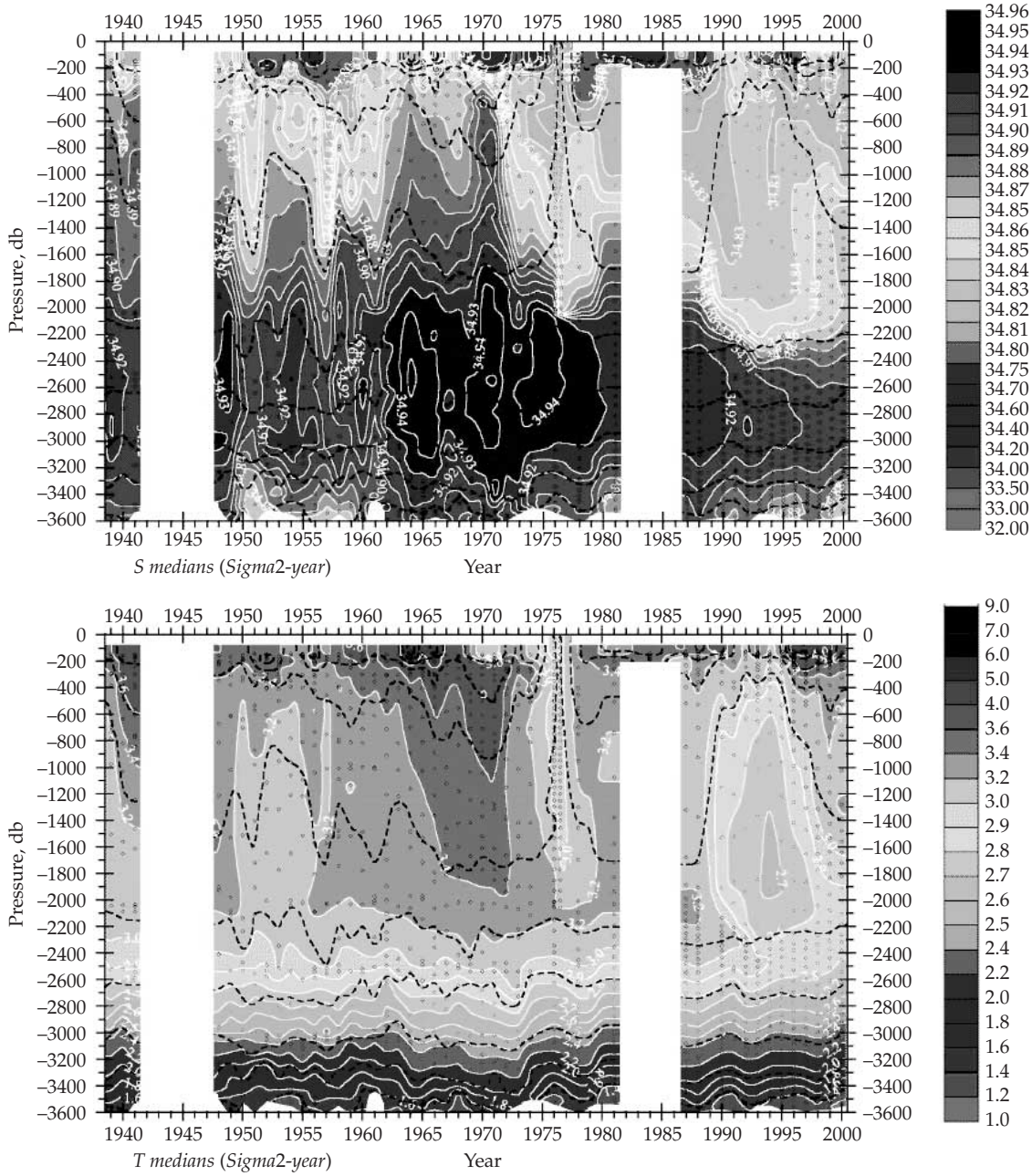

Figure 2.9 Changes in the potential temperature (θ; upper panel) and salinity (lower panel) of the watercolumn in the Central Labrador Sea over the complete period of the hydrographic record since 1938. The data set was selected to lie within the 3300-m isobath of the Labrador Sea, and the plots represent the median values of vertical property profiles, binned according to σ_2 density intervals. Kindly provided by Igor Yashayaev. (See Plate 3.)

pycnostad (Jenkins 1982; Talley and Raymer 1982; Talley 1996). With winter storm activity concentrated at the US eastern seaboard, storminess decreased to a postwar minimum further north so that Labrador Sea convection became increasingly suppressed and freshwater built up at the surface (Lazier 1980, 1988, 1995).

The tendency towards opposite conditions prevailed over the subsequent 20 years (Fig. 2.9; see also Dickson *et al*. 1996; Dickson *et al*. 2000), so that by the early 1990s, Labrador Sea Water (LSW) was fresher, colder, denser, and deeper than at any other time in the history of deep measurements in the region. From 1966 to 1994, spanning most of the period of the trend towards more positive NAO index winters, the overall freshening of the water column of the Labrador Sea was equivalent to mixing down an extra 7 m of freshwater from the sea surface, and its cooling was equivalent to an increased loss rate of 8 W m^{-2} (Lazier 1995). These are among the largest changes observed in oceanography.

Such altered water mass properties have value in tracing-out the rates and pathways by which LSW spreads across the basin. It is this cold, fresh, and dense new vintage of LSW that Sy *et al*. (1997) use, together with its chlorofluorocarbon signature, to derive modern estimates of LSW spreading rates that are an order of magnitude greater than published values within the Labrador-Irminger Basin, and perhaps three to four times greater than previous trans-ocean estimates (e.g. Read and Gould 1992). The influence of these changes on the Atlantic gyre circulation itself, however, is likely to be of even greater importance.

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The main North Atlantic Current is driven by the gradient of potential energy anomaly (PE) across the mutual boundary between the subtropical and subpolar gyres (Curry and McCartney 2001). Since PE reflects the vertical density structure and heat content of the upper ocean to well below the wind-driven layer, coordinated changes of opposite sign in the production and characteristics of the mode waters in each gyre will have the potential to drive deep-seated changes in the PE gradient and, hence, in the strength of the Atlantic gyre circulation. If these changes in the density and heat content of mode waters are attributable to NAO forcing, moreover, the amplification of the NAO to extreme values over recent decades is likely to have been accompanied by a corresponding multi-decadal spin-up of the Atlantic gyre circulation. Specifically, Curry and McCartney (2001) calculate a 30% increase to the mid-1990s in the 0–2000 dbar east-going baroclinic mass transport along the gyre boundary, with both gyres contributing equally. Moreover, if the link to NAO forcing proves valid, probably at no other time over the twentieth century did the North Atlantic gyre circulation exceed its strength during the early 1990s.

2.3.4 Indirect response to NAO forcing at abyssal depths

Below the LSW layer in the Labrador Sea at depths of 2300–3500 m, repeat hydrography has indicated a steady freshening over the past three to four decades (Fig. 2.9(a)). At these depths, beyond the reach of deep convection, such a change cannot be due to local forcing. There is growing evidence that, instead, it reflects a large-scale freshening of the upper Nordic Seas passed on via the dense northern overflows that cross the Greenland–Scotland Ridge through the Denmark Strait and the Faroe Bank Channel. The freshening of the Arctic and subarctic seas is, however, itself associated with the multi-decadal NAO variability that forms the focus of this chapter. As such, the changes at abyssal depths may be viewed as an indirect response to atmospheric forcing. The factors outlined below provide an impressive illustration of the varied mechanisms and locations by which a change in NAO forcing may translate into a large-scale ocean response.

1. The direct export of sea-ice from the Arctic Ocean is one such cause of recent subarctic freshening. A combination of current measurements, upward-looking sonar and satellite imagery reveals an increased annual efflux of ice through the western Fram Strait to a record volume-flux of 4687 km^3 $year^{-1}$ in 1994–95 (Vinje *et al*. 1998; Kwok and Rothrock 1999). Although the relationship is not robust in the longer-term, each 1Σ increase in the NAO index since 1976 has been associated with an approximately 200 km^3 increase in the annual efflux of ice (Dickson *et al*. 2000).

2. Throughout the marginal ice-zone of the Nordic Seas, a steady decrease in the local winter production of sea-ice at the end of winter has been associated with the trend in the NAO over the past 40 years (Deser *et al*. 2000).

3. Precipitation along the Norwegian Atlantic Current increases during NAO-positive conditions

by approximately 15 cm per winter compared with NAO-negative conditions, associated with the aforementioned extension of storm activity to the Nordic Seas (Dickson *et al*. 2000).

Other factors and mechanisms have undoubtedly contributed to the long and gradual but dramatic freshening of the European subarctic seas in recent decades, most of them associated in some way with the amplifying NAO. Blindheim *et al*. (2000) describe a range of factors internal to the Nordic Seas. These include an increased freshwater supply from the East Icelandic Current, a narrowing of the salty Norwegian Atlantic Current towards the Norwegian Coast, a dual change towards reduced deep water formation and increased Arctic Intermediate Water formation in the Greenland Sea (Dickson *et al*. 1996; Verduin and Quadfasel 1999) that has had the effect of deepening the interface between the Arctic Intermediate Water and Deep Water in the Norwegian Sea. Other more-remote influences may even include an observed freshening of the Pacific inflow to the Arctic Ocean through the Bering Strait (Carmack 2000, personal communication), and a contribution from an apparent recent thinning of sea-ice in the Arctic Ocean (Rothrock *et al*. 1999). In turn, the warming and perhaps strengthening of Atlantic water inflow to the Arctic Ocean due to the amplifying NAO (Quadfasel *et al*. 1991; Grotefendt *et al*. 1998; Dickson *et al*. 2000), may itself have contributed to ice-thinning in the Eurasian Basin of the Arctic Ocean (Steele and Boyd 1998); by the late 1980s and early 1990s, both inflow streams were between 1 and 2 °C warmer than normal with a consequent warming, spreading, and shoaling of the Atlantic-derived sub-layer across the Eurasian Basin of the Arctic Ocean (Carmack *et al*. 1995; Aagaard *et al*. 1996; Carmack *et al*. 1997; Swift *et al*. 1997; Morison *et al*. 1998a,b; Morison *et al*. 2001).

While it may not be possible to partition the recent freshening of the Nordic Seas into its individual contributory components, it is clear from three of the longest observational records available that the climate change has occurred over a sufficiently deep layer (approximately 1–1.5 km) to affect the hydrographic character of the two dense overflows, which cross the Greenland–Scotland Ridge and ventilate the abyssal North Atlantic, thus helping to drive the MOC. Hydrographic sections monitoring the outflow of North East Atlantic Deep Water (NEADW) through the Faroe–Shetland Channel confirm that NEADW salinities have decreased linearly by ≈0.01 per decade since the mid-1970s (Turrell *et al*. 1999; Dickson *et al*. 2001), precisely the rate, period, and steadiness of freshening recorded downstream in the NEADW layer of the abyssal Labrador Sea. A similar rate and period of freshening is also found in the deepest layers of the Labrador Sea, which is occupied by water that overflows the Greenland–Scotland Ridge via the Denmark Strait (DSOW). Both dense overflows, therefore, appear to have tapped and delivered to the headwaters of the THC the freshening signal of the upper Nordic Seas (Fig. 2.10). In addition, near-surface variability in temperature generated by the NAO in the eastern Fram Strait appears to re-circulate to the south, affecting the temperature of the deepening dense overflow of DSOW off of southeast Greenland and from there determining the density of abyssal depths of the Labrador Sea (Dickson *et al*. 1999, 2001).

The subarctic seas are thus observed to act in two capacities: first, as a source of change for the climatically-sensitive Arctic Ocean, and second as a mechanism for transferring such changes to the MOC. Directly, or indirectly, the signature of the trend in NAO toward its positive index state in recent decades can be identified in Atlantic Ocean variability over the full range of ocean depths.

2.3.5 Space- and time dependence in NAO forcing and ocean response

The climate record is noisy, reflecting a continuum of variability on all space and timescales. As Wunsch (1992) points out 'The ocean is a turbulent fluid in intimate contact with another turbulent fluid, the atmosphere. Although I am unaware of any formal theorems on the subject, experience with turbulent systems suggests that it is very unlikely that any components of such a complex nonlinear system can actually remain fully steady; . . . the frequency/wavenumber spectrum of the ocean circulation is almost surely everywhere filled.' As described in the section '*Atypical NAO winters and other atmospheric variations*', the chaotic nature of the atmospheric circulation means that significant departures from the anomalies depicted schematically in Fig. 2.3 occur more often than not. Since the NAO is itself variable in both space and time, so too is the ocean (and ecosystem) response.

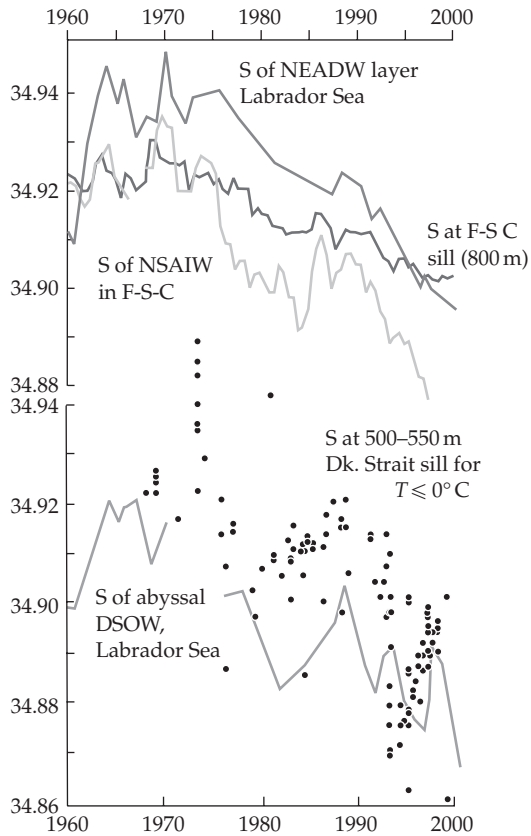

Figure 2.10 Evidence that the recent freshening of the upper Nordic Seas is passing across the Greenland–Scotland Ridge via both intermediate-depth overflows to affect the deep and abyssal layers of the North Atlantic. The upper group of curves indicate a freshening by ≈0.01 per decade at sill depth in the Faroe–Shetland Channel and in the Norwegian Sea Arctic Intermediate Water layer (NSAIW) overflowing the Faroe–Shetland Channel, together with a corresponding decrease downstream in the NEADW of the Labrador Sea. The lower curves show that a very similar rate of freshening has been observed at sill depth in the Denmark Strait (dots) and in the DSOW layer that occupies abyssal depths in the Labrador Sea. From Dickson *et al*. (2001).

Changes in the NAO amplitude over time

While the recent amplification of the NAO has brought a progressive warming to parts of the subpolar gyre and Barents Sea (e.g. Dickson *et al*. 2000, their figure 10), the NAO does not appear to have been the dominant cause of warming in this region during the middle decades of the twentieth century. Between the 1920s and 1960s, the correlation between the NAO index and the temperature of the Barents Sea dropped to a long-term minimum (Dickson *et al*. 2000, their figure 14) and instead, abnormally warm and saline conditions appear to have passed through subpolar gyre from other causes, bringing an amelioration of the marine climate to northern waters from the West Greenland Banks to Iceland, Faroes, and the Barents Sea (Dickson and Brander 1993). It was not until the cold, fresh conditions of the GSA passed through the northern gyre in the late 1960s and 1970s (Dickson *et al*. 1988; Reverdin *et al*. 1997), that this warming came to an abrupt end.

The mechanisms and processes responsible for the so-called 'Warming in the North' are still under investigation (e.g. Delworth and Knutson 2000). Its signature, as well as its impact on the marine ecology, however, is unmistakable:

- The salinity of North Atlantic Water passing through the Faroe–Shetland Channel reached a century-long high (Dooley *et al*. 1984).
- Salinities were so high off Cape Farewell that they were thrown out as erroneous (Harvey 1962).
- A precipitous warming of more than 2 °C in the 5-year mean pervaded the West Greenland Banks (Dickson and Brander 1993).
- The northward dislocations of bio-geographical boundaries for a wide range of species from plankton to commercial fish, terrestrial mammals and birds were at their most extreme in the twentieth century.

The astonishing nature of these radical events is vivid in the contemporary scientific literature, as summarized in a comprehensive bibliography by Lee (1949) and reviewed in an ICES Special Scientific Meeting on 'Climatic Changes in the Arctic in Relation to Plants and Animals' in 1948. The rise and spread of the West Greenland cod fishery was one spectacular result, apparently reflecting a change in the effectiveness of egg and larval drift in the Irminger/West Greenland Current system (Buch and Hansen 1988; Dickson and Brander 1993; Schopka 1993, 1994). While similar 'cod periods' at West Greenland also occurred in the 1820s and 1840s (Hansen 1949), and perhaps much earlier (Fabricius 1780), the change in the middle decades of the twentieth century represented a return of cod to the region after a 50–70 year absence (Buch and Hansen 1988; Dickson *et al*. 1994), only to end again with the arrival of the GSA.

Radical changes in the Atlantic ecosystem were not confined to W. Greenland and the Davis Strait in these middle decades of the Century. The

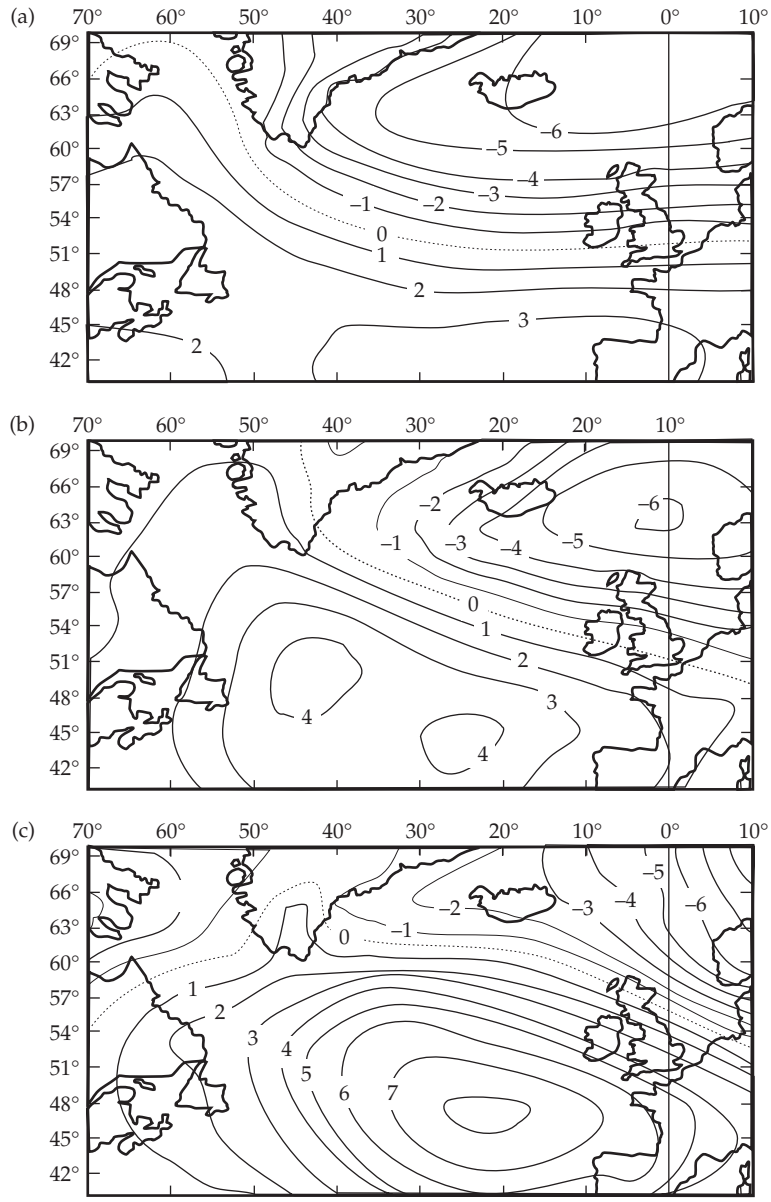

Figure 2.11 SLP anomaly pattern for winters (a) 1993–95, (b) 1999, and (c) 2000.

'Russell Cycle' brought a more-southerly community to the ecosystem of the Western Channel at about the same time and perhaps for related reasons (e.g. Cushing 1982). In the Nordic Seas, the spawning stock biomass of the Norwegian spring-spawning herring rose to a maximum in mid-Century as the wave of warmth passed through the northern Gyre (Toresen and Ostvedt 2000), and the range of its feeding-spawning-overwintering movements expanded to the west and north following the retreat of the Ocean Polar Front (since reversed; see Vilhjalmsson 1997). In the Barents Sea the wave of warmth brought a sustained increase in the yield, weight, liver-weight, roe weight, and

recruitment of *skrei* from their pre-existing minima associated with the extreme cold in the early years of the Century (Helland-Hansen and Nansen 1909; Anon 1996).

2.3.6 Temporal shifts in the NAO pattern

As described earlier, NAO-positive winters are usually associated with chill, dry northwesterly winds across the Labrador Sea, and hence intense and deep-reaching convective renewal of LSW and a widespread distribution of chilled surface temperatures across the northwest Atlantic (Fig. 2.6(b)). For instance, correlations between the winter NAO index and both temperature and salinity on top of Fylla Bank, West Greenland, are strongly negative (table 10.1 of Buch 1995). However, this relationship has failed to hold following the temporary deep minimum of the NAO index in the winter of 1996 (Fig. 2.2). The NAO-positive conditions of the most recent winters have been accompanied by a shutdown of Labrador Sea convection and the warmest conditions on top of Fylla Bank evident in the 50-year record (Buch 2000). In just two or three winters, the long sustained cooling and freshening of LSW (Fig. 2.9), described in Section 2.3.4. as 'among the largest changes observed in oceanography', has been largely reversed. The suggested cause is shown in Fig. 2.11. Comparing the Atlantic SLP anomaly pattern for the extreme NAO-positive winters of 1993–95 with that for 1999–2000 reveals a slight east or northeast displacement in the more recent period. This subtle shift has little effect along the eastern boundary to the Barents Sea where widespread warming has continued. In the northwest Atlantic, however, the slight eastward displacement of the 'normal' NAO pattern has made an important difference to the marine climate of the West Greenland Banks and to the convective centre of the Labrador Sea. Instead of a chilling northwesterly flow of air, light anomalous southerly winds have prevailed across the Labrador Sea to West Greenland in 1999–2000.

The salient point is that the interannual 'noise' of the extratropical atmosphere is large, and even subtle shifts of recurrent climate patterns such as the NAO may have deep-reaching and long-lasting effects. Moreover, to the extent that anthropogenic climate change might influence modes of natural variability, perhaps making it more likely that one phase of the NAO is preferred over the other (e.g. Corti *et al*. 1999; Gillett *et al*. 2000, 2001), or that subtle shifts might occur in its spatial pattern (e.g. Ulbrich and Christoph 1999), it would be unwise to rely entirely on past experience in defining the 'typical' Atlantic Ocean (and ecosystem) response to NAO forcing. Schematics such as Fig. 2.3 should therefore be used with caution!