DECADAL VARIATIONS IN CLIMATE ASSOCIATED WITH THE NORTH ATLANTIC OSCILLATION

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Abstract. Large changes in the wintertime atmospheric circulation have occurred over the past two decades over the ocean basins of the Northern Hemisphere, and these changes have had a profound effect on regional distributions of surface temperature and precipitation. The changes over the North Pacific have been well documented and have contributed to increases in temperatures across Alaska and much of western North America and to decreases in sea surface temperatures over the central North Pacific. The variations over the North Atlantic are related to changes in the North Atlantic Oscillation (NAO). Over the past 130 years, the NAO has exhibited considerable variability at quasi-biennial and quasi-decadal time scales, and the latter have become especially pronounced the second half of this century. Since 1980, the NAO has tended to remain in one extreme phase and has accounted for a substantial part of the observed wintertime surface warming over Europe and downstream over Eurasia and cooling in the northwest Atlantic. Anomalies in precipitation, including dry wintertime conditions over southern Europe and the Mediterranean and wetter-than-normal conditions over northern Europe and Scandinavia since 1980, are also linked to the behavior of the NAO. Changes in the monthly mean flow over the Atlantic are accompanied by a northward shift in the storm tracks and associated synoptic eddy activity, and these changes help to reinforce and maintain the anomalous mean circulation in the upper troposphere. It is important that studies of trends in local climate records, such as those from high elevation sites, recognize the presence of strong regional patterns of change associated with phenomena like the NAO.

1. Introduction

An objective of the International Workshop on Climatic Change at High Elevation Sites (HIGHEST-95) was to highlight the value of high elevation geophysical records and to evaluate climatic trends that may be present in the records. The pristine and remote nature of many high elevation sites may make their data collections especially valuable for the detection of a human influence on climate. Paleoclimatological and historical evidence show that mountain areas are highly-sensitive to changes in climate (e.g., Barry, 1990), and the retreat of glaciers over the past century in many parts of the world (Haeberli et al., 1989; Oerlemans, 1994) is broadly consistent with the observed rise in global mean surface temperatures of 0.3° to 0.6 °C over the same period (IPCC, 1996). Such long-term trends, however, are not uniform and exihibit considerable spatial and short-term variability. For instance, twentieth century ice loss is not evident in the glaciers of the Canadian Arctic (Fisher and Koerner, 1994), and in recent years the maritime glaciers of southwest Norway have exhibited positive mass balances (World Glacier Monitoring Service,

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1994). Similarly, global temperatures decreased from the late 1930s to the mid-1970s, but since the mid-1970s temperatures have increased to record levels during the past decade (IPCC, 1996). The recent warming has been largest during the winter and spring seasons over the Northern Hemisphere (NH) continents between 40° N and 70° N, while cooling has occurred over the northern oceans (Figure 1a).

The purpose of our paper at HIGHEST-95 was to illustrate that the recent pattern of NH temperature change is strongly related to decade-long changes in the large-scale circulation of the atmosphere and ocean (Palecki and Leathers, 1993; Wallace et al., 1995, 1996; Hurrell, 1996). The observed presence of strong regional patterns of change must be recognized in interpreting local climate records, such as those from high elevation sites discussed in other papers of this issue of *Climatic Change*. We show that the recent cooling over the northwest Atlantic and the warming across Europe and downstream over Eurasia since the early 1980s (Figure 1a) is directly related to decadal changes in the North Atlantic Oscillation (NAO), while the temperature anomalies over the Pacific basin and North America result in part from tropical forcing associated with the El Niño-Southern Oscillation (ENSO) phenomenon but with important feedbacks in the extratropics. The changes in circulation over the Atlantic are also linked to coherent large-scale anomalies in precipitation since the early 1980s including dry conditions over southern Europe and the Mediterranean and wetter-than-normal conditions over northern Europe and parts of Scandinavia. The results summarize and expand upon the findings of Hurrell (1995a, 1996) and emphasize the point that the NAO, in addition to the Southern Oscillation (SO), is a major source of interannual variability of weather and climate around the world.

2. Changes in the Pacific

Decadal variations in the climate over the North Pacific with associated teleconnections downstream across North America have long been of interest and have been particularly highlighted by the work of Namias (1959, 1963, 1969); see also Dickson and Namias (1976), Douglas et al. (1982) and Namias et al. (1988). Recently, a large amount of evidence has emerged of a substantial change in the wintertime atmospheric circulation over the North Pacific that began in the mid-1970s and lasted throughout the 1980s. The changes involved the Pacific-North American (PNA) teleconnection pattern and corresponded to a deeper and eastward shifted Aleutian low pressure system (Figure 1b) which advected warmer and moister air along the west coast of North America and cooler and drier air over the central North Pacific (Nitta and Yamada, 1989; Trenberth, 1990; Trenberth and Hurrell, 1994). Consequently, there were increases in temperatures and sea surface temperatures (SSTs) along the west coast of North America and Alaska but decreases in SSTs over the central North Pacific (Figure 1a). Changes in coastal rainfall and streamflow have also been noted (Cayan and Peterson, 1989), as well as decreases

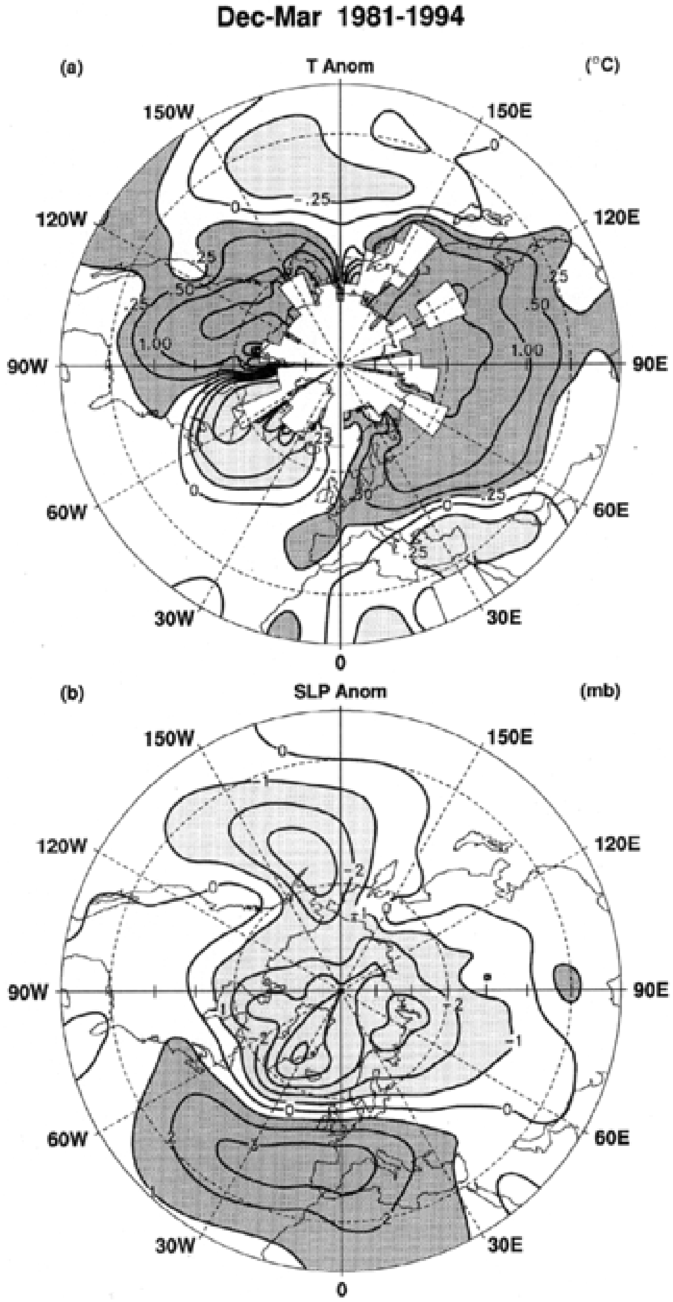

Figure 1. Fourteen winter (1981–1994) average (a) surface temperature and SST anomalies and (b) SLP anomalies expressed as departures from the 1951–1980 means. Temperature anomalies $> 0.25\,^{\circ}$C are indicated by dark shading, and those $< -0.25\,^{\circ}$C are indicated by light shading. The same shading convention is used for SLP anomalies greater than 1 mb in magnitude. Regions of insufficient data coverage are not contoured in (a).

in sea ice in the Bering Sea (Manak and Mysak, 1987). Changes in the mean flow were accompanied by a southward shift in the storm tracks and associated synoptic eddy activity (Trenberth and Hurrell, 1994) and in surface ocean sensible and latent heat fluxes (Cayan, 1992). On decadal and longer time scales, changes in the ocean must also become a factor in maintaining the extratropical circulation anomalies (Latif and Barnett, 1994).

A simple index used to measure the variations over the North Pacific (NP) is the area-weighted mean sea level pressure (SLP) over the region 30° to 65° N, 160° E to 140° W (Trenberth and Hurrell, 1994). The NP index (for North Pacific), averaged over the winter months from December through March, is shown in Figure 2 since 1925 (the given year corresponds to the January of the winter season). Pressures from 1977 to 1988 were lower by 2.2 mb relative to the 70-winter NP-area mean. The only previous period that comparable values occurred was for a much shorter interval in the early 1940s. The pattern of temperature change associated with the NP index (Figure 3) shows that the recent North Pacific basin anomalies (Figure 1a) are consistent with the longer record: below normal NP values are associated with below-normal temperatures over the North Pacific and above normal surface temperatures along the west coast of North America extending into Alaska and across much of Canada. The departure pattern during winter also reveals below-normal temperatures over the southeast United States, which illustrates the PNA teleconnection and occurs in opposition to the temperature changes associated with NAO (see Figure 6).

The decadal changes over the North Pacific have been linked to variations in the tropics (Trenberth and Hurrell, 1994; Kawamura, 1994), and several modeling studies have confirmed that North Pacific atmospheric variability is controlled in part by anomalous tropical Pacific SST forcing (Kitoh, 1991; Chen et al., 1992; Lau and Nath, 1994; Graham et al., 1994; Miller et al., 1994; Kumar et al., 1994). Fluctuations in tropical SSTs are related to changes in ENSO, and the observed warming of the tropical waters since the mid-1970s (e.g., Trenberth and Hoar, 1996) has been linked to increased tropospheric temperatures and water vapor in the western Pacific (Hense et al., 1988; Gaffen et al., 1991; Gutzler, 1996) and a more active hydrological cycle (Nitta and Yamada, 1989; Graham, 1995). The variability of the SO is evident in the NP index (Figure 2), but feedback effects in the extratropics may serve to emphasize the decadal over interannual time scales relative to the tropics (Trenberth and Hurrell, 1994).

3. Changes in the Atlantic

a. The NAO Index

Over much of the past decade, anomalous coldness during winter has prevailed near Greenland and the eastern Mediterranean, while very warm conditions have been notable over Scandinavia, northern Europe, the former Soviet Union and

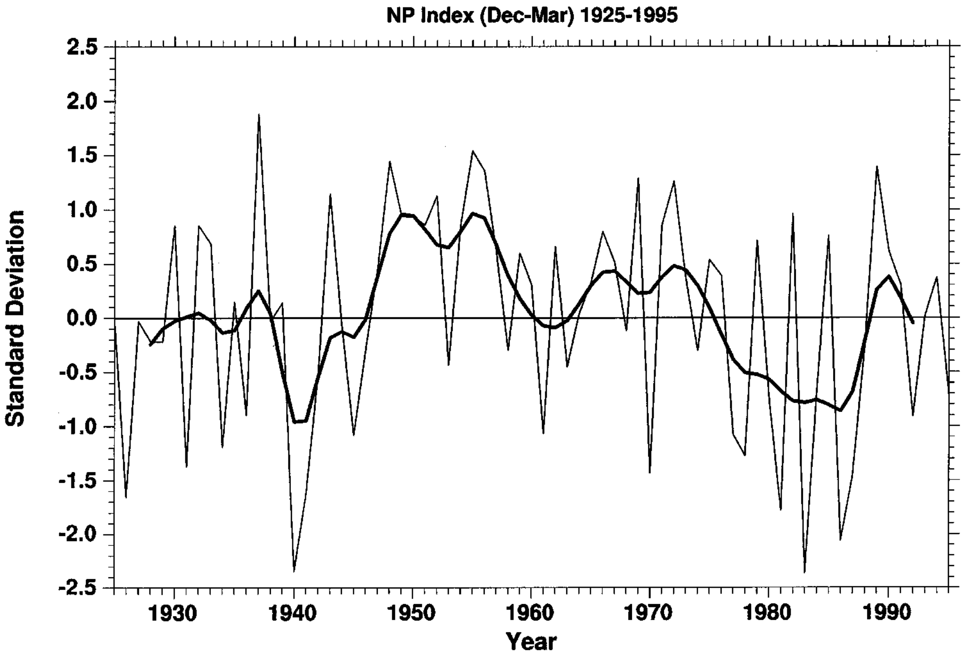

Figure 2. Time series of normalized mean North Pacific SLP averaged over 30° to 65° N, 160° E to 140° W (the NP index) for the months December–March beginning in 1925 and smoothed with a low-pass filter with seven weights to remove fluctuations with periods less than 4 years.

much of central Asia (Figure 1a). Interannual and longer time-scale changes in the atmospheric circulation and lower tropospheric temperatures during winter over the North Atlantic and adjacent land areas do not appear to be strongly influenced by tropical SST variability (e.g., Barnett, 1985; Kumar et al., 1994; Graham et al., 1994; Lau and Nath, 1994). Rather, the anomalies are more strongly linked to the recent behavior of the NAO.

The NAO, which is associated with changes in the surface westerlies across the Atlantic onto Europe, refers to a meridional oscillation in atmospheric mass with centers of action near the Icelandic low and the Azores high (e.g., van Loon and Rogers, 1978). Although it is evident throughout the year, it is most pronounced during winter and accounts for more than one-third of the total variance of the SLP field over the North Atlantic (Figure 4, see also Wallace and Gutzler, 1981; Barnston and Livezey, 1987). Because the signature of the NAO is strongly regional, a simple index of the NAO can be defined as the difference between the normalized mean winter (December–March) SLP anomalies at Lisbon, Portugal and Stykkisholmur, Iceland (Hurrell, 1995a). The winter SLP anomalies at each station were normalized by dividing each seasonal pressure by the long-term mean (1964–1995) standard deviation. The variability of the NAO index since 1864 is shown in Figure 5, where the heavy solid line represents the low pass filtered meridional pressure gradient.

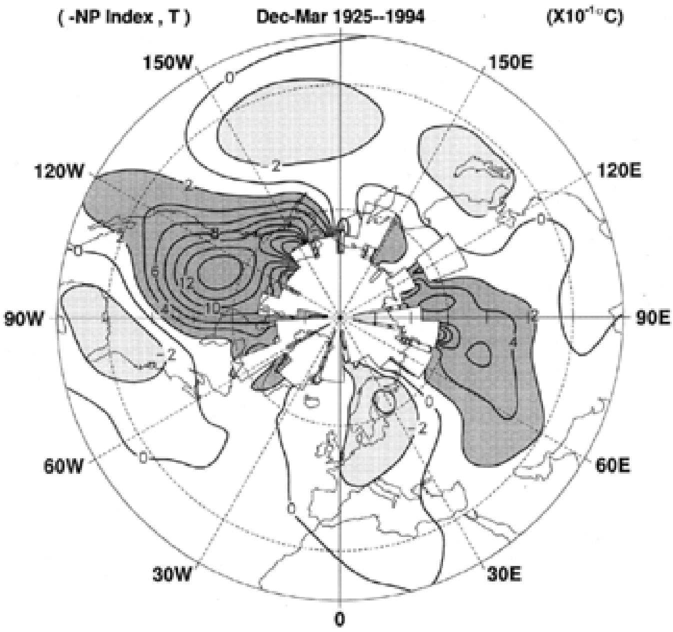

Figure 3. Changes in temperatures ($\times 10^{-1}$ °C) corresponding to a unit deviation of the NP index, multiplied by minus one, computed over the winters from 1925 through 1994. The contour increment is 0.2 °C. Temperature departures > 0.2 °C are indicated by dark shading, and those < -0.2 °C are indicated by light shading. Regions of insufficient data are not contoured.

Positive values of the index indicate stronger-than-average westerlies over the middle latitudes associated with low pressure anomalies over the region of the Icelandic low and anomalously high pressures across the subtropical Atlantic.

In addition to a large amount of interannual variability, there have been several periods when the NAO index persisted in one phase over many winters (van Loon and Rogers, 1978; Barnett, 1985). Over the region of the Icelandic low, seasonal pressures were anomalously low during winter from the turn of the century until about 1930 (with the exception of the 1916–1919 winters), while pressures were higher than average at lower latitudes. Consequently, the wind onto Europe had a strong westerly component, and the moderating influence of the ocean contributed to higher than normal temperatures over much of Europe (e.g., Parker and Folland, 1988). From the early 1940s until the early 1970s, the NAO index exhibited a downward trend, and this period was marked by European wintertime temperatures that were frequently lower than normal (van Loon and Williams, 1976; Moses et al., 1987). A sharp reversal has occurred over the past 25 years and, since 1980, the NAO has remained in a highly-positive phase with SLP anomalies of more than 3 mb in magnitude over both the subpolar and subtropical Atlantic (Figure 1b, see also Walsh et al., 1996). The 1983, 1989 and 1990 winters were marked by the highest positive values of the NAO index recorded since 1864 (Figure 5). Beniston

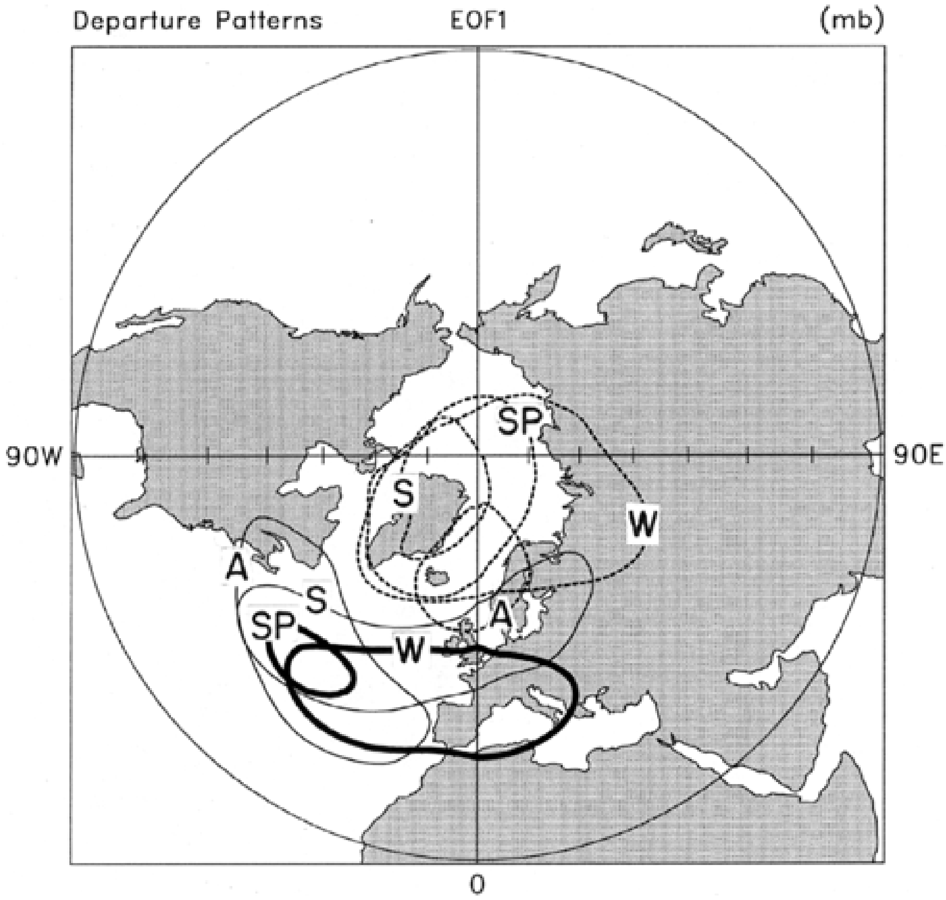

Figure 4. The first empirical orthogonal function (EOF) of unnormalized SLP based on data over the region 20° to 80° N, 90° W to 40° E from 1899–1994. The values represent the changes in SLP (mb) over the hemisphere associated with a unit deviation of the time series of the first principal component (EOF1). Values are shown for each of four seasons, winter (W, December–February), spring (SP, March–May), summer (S, June–August), and autumn (A, September–November). SLP departures of –2 mb are indicated by the dashed lines, departures of 0.5 mb are indicated by the thin solid lines, and departures of 2 mb are indicated by dark, heavy lines.

et al. (1994) note that blocking highs over Switzerland were more frequent during the 1980s than at any other time this century, with a decadal frequency 2–3 times greater than during previous decades. Nearly 25% of the total observed blocking highs over Switzerland since 1900 occurred during the 1980s.

The NAO index in Figure 5 differs from that of Rogers (1984) who, simplifying the more complicated index of Walker and Bliss (1932), defined the NAO index using sea level pressure anomalies from Ponta Delgada, Azores and Akureyri, Iceland. The record at Ponta Delgada available from the World Monthly Surface Station Climatology begins in 1894, however, so Hurrell (1995a) selected Lisbon

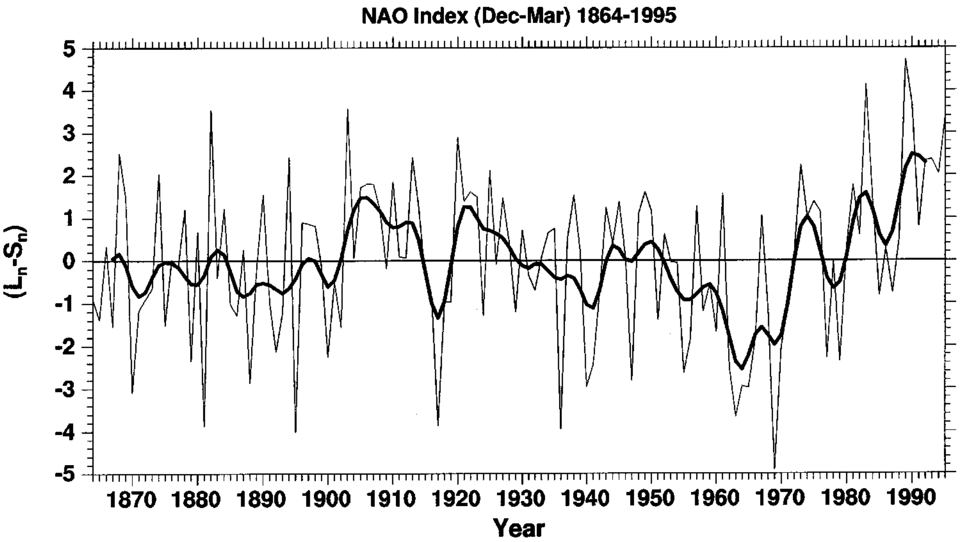

Figure 5. Time series of the winter (December-March) index of the NAO (as defined in the text) from 1864–1995. The heavy solid line represents the meridional pressure gradient smoothed with a low pass filter to remove fluctuations with periods less than 4 years.

and Stykkisholmur in order to extend the record an additional 30 years. The correlation between the NAO index in Figure 5 and the index of Rogers (1984) over the years 1894–1995 is 0.93. Since an NAO index based on station pressures will be affected by small-scale and transient meteorological phenomena not related to the NAO, the index will contain noise. Following Trenberth (1984), the signal-to-noise ratio of the NAO index can be assessed through the simple formula

$$\frac{\text{Signal}}{\text{Noise}} = \left[\frac{1 - r_{SL}}{1 + r_{SL}}\right]^{1/2} \tag{1}$$

where r_{SL} (or r_{SD}) is the cross correlation between seasonal (3 month) pressure anomalies at Stykkisholmur (S) and Lisbon (L) or Ponta Delgada (D). The noise, in this case, is a measure of all fluctuations where the two centers of action in the NAO are operating in phase and therefore are not part of the oscillation. The use of Stykkisholmur in place of Akureyri does not significantly affect the results as the December–March anomalies at these two Icelandic stations correlate at 0.98. The seasonal cross correlations and signal-to-noise ratios in Table I confirm the patterns in Figure 4. During northern winter, the subtropical node of the NAO is well-captured by either Lisbon or Ponta Delgada, although the signal-to-noise ratio is slightly higher when Lisbon is used in the index. The westward migration of the subtropical high through spring and summer, however, dictates that Ponta Delgada is the better station to use for the other seasons or for an annual NAO index. For the December–March season, the signal-to-noise ratio of the Stykkisholmur-Lisbon (Ponta Delgada) index is 2.6 (2.2).

Table I

Correlation coefficients r between seasonal SLP anomalies over the years 1894–1995 between Stykkisholmur and Lisbon ($S,\ L$) and Stykkisholmur and Ponta Delgada ($S,\ D$). Also given are the signal-to-noise ratios (S/N) for each season according to (1)

Season	$r(S,\ L)$	S/N	$r(S,\ D)$	S/N
DJF	–0.68	2.3	–0.61	2.0
JFM	–0.75	2.6	–0.73	2.5
FMA	–0.64	2.1	–0.69	2.3
MAM	–0.35	1.4	–0.58	1.9
AMJ	–0.06	1.1	–0.35	1.4
MJJ	0.05	1.0	–0.30	1.4
JJA	0.13	0.9	–0.29	1.3
JAS	0.03	1.0	–0.36	1.5
ASO	–0.11	1.1	–0.48	1.7
SON	–0.22	1.3	–0.46	1.6
OND	–0.40	1.5	–0.47	1.7
NDJ	–0.50	1.7	–0.51	1.8

b. Relationships to Temperature

The changes in local surface temperatures and SSTs based on linear regression with the NAO index are shown in Figure 6. The surface temperature data are the same as those used in Figure 1a and consist of land surface temperatures blended with SST data (Jones and Briffa, 1992; Parker et al., 1994). Changes of more than 1 °C associated with a one standard deviation change in the NAO index occur over the northwest Atlantic and extend from northern Europe across much of Eurasia. Changes in temperatures over northern Africa and the southeast United States are also notable. The similarity between the departure pattern of temperature (Figure 6) and the decadal anomalies over the North Atlantic and surrounding landmasses (Figure 1a) is striking and suggests that the recent temperature anomalies over these regions are strongly related to the persistent and exceptionally strong positive phase of the NAO index since the early 1980s.

The effect of circulation changes on temperature can be quantified through multivariate linear regression (Palecki and Leathers, 1993). Previously, a common application of linear regression has been to remove the influence of the SO from hemispheric and global temperature time series (e.g., Jones, 1994; Christy and McNider, 1994). Hurrell (1996) regressed the both the NAO index and an index of the SO upon the NH extratropical (20° N to 90° N) temperature anomalies for each winter since 1935, and found that 44% of the variance of the temperatures could be explained. Variations of temperature associated with the NAO account for 31% of the hemispheric interannual variance, while the SO accounts for 16% (see his

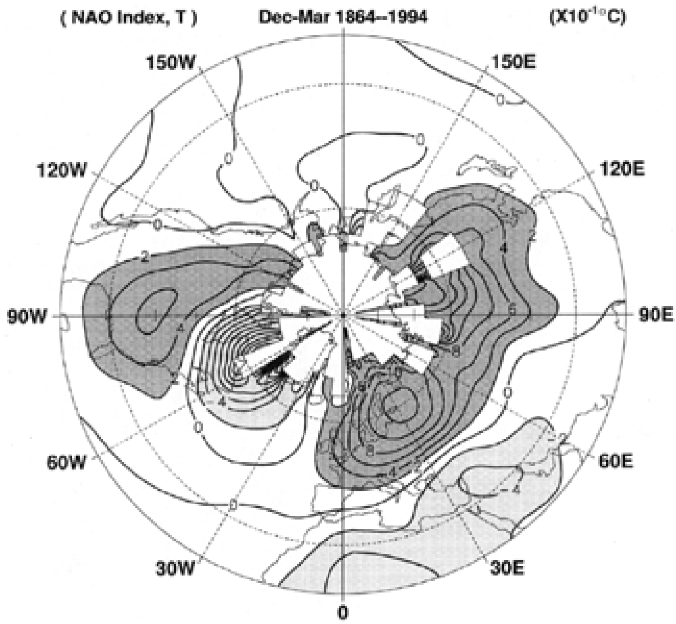

Figure 6. As in Figure 3, but for changes in temperatures ($\times 10^{-1}$ °C) corresponding to a unit deviation of the NAO index computed over the winters from 1864 through 1994.

Table I). Moreover, the NAO and the SO account linearly for much, but not all, of the hemispheric warming since the early 1980s. Over the period 1981 to 1994, the NH extratropical temperature anomaly relative to the 1935–1994 winter mean was 0.29 °C, but after the removal of the linear temperature contributions by the NAO and the SO, the residual warming was 0.07 °C. Of the 0.22 °C difference, the NAO-related warming was 0.15 °C. Locally, nearly all of the cooling in the northwest Atlantic and the warming across Europe and downstream over Eurasia since 1980 can be linearly related to changes in the NAO (see Figure 5 of Hurrell, 1996).

c. Time Scales

To explore which time scales contribute to the strong relationship between the NAO and temperatures across the Atlantic basin, the relationship between the NAO index and winter temperatures over northern Europe can be examined as a function of frequency. Figure 7 shows the power spectrum of the NAO index for the 130 winters 1865–1994. The spectrum is scaled so that the sum of the spectral estimates is the same as the temporal variance of the NAO index. Also shown is the corresponding red noise spectrum with the same lag one autocorrelation coefficient and the 5% and 95% confidence limits. The spectrum reveals significant variance at biennial periods, a deficit in power at 3 to 5 year periods, and enhanced power at 6 to 10 year periods. The power at the lowest frequencies reflects the trends evident in

Figure 5. The time evolution of these signals is shown as a contour plot of the power spectra computed from running 60 year intervals beginning with the time period 1865–1924 and ending with 1935–1994 (Figure 7). Much of the variance at biennial periods comes from early in the record, while the variability between 6 and 10 years is present throughout the record but has become most pronounced over the latter half of this century. The tendency for the NAO power spectrum to become redder with time is clearly indicated by the running 60-year lag one autocorrelations also shown in Figure 7.

The power spectrum of winter (December–March) temperature anomalies from Copenhagen, Denmark over the same period (Figure 8) is very similar in character to that of the NAO index. The spectra of Copenhagen temperatures are dominated by variations at biennial periods early in the record, and this frequency band accounts for a significant percentage of the total variance throughout the 130-year record. The variance at 6 to 10 year periods is enhanced later in the data record and accounts for the largest percentage of the total variance over the last several 60 year periods. The spectra in Figures 7 and 8 suggest a strong relationship between the NAO index and temperatures over northern Europe, but one that has changed over time. This nonstationary behavior is documented in Figure 9, which shows the coherence squared between the NAO and temperature anomalies at Copenhagen as a function of time and frequency. Correlation coefficients between the NAO index and wintertime temperatures over northern Europe are generally > 0.7 (not shown). These correlations result from coherent fluctuations at biennial periods early in the data records and lower-frequency fluctuations at 6 to 10 year periods later on (Figure 9).

d. Changes in Storm Tracks and Their Effects on the Mean Flow

Changes in the mean circulation patterns over the North Atlantic are accompanied by pronounced shifts in the storm tracks and associated synoptic eddy activity (Rogers, 1990; Hurrell, 1995b; Rogers and Mosley-Thompson, 1995). To provide an analogy to the decadal changes evident in the NAO index (Figure 5), low or near normal NAO index winters are compared with very-high NAO index winters using composites of the global analyses produced by the European Centre for Medium Range Weather Forecasts (ECMWF). For the low or normal NAO index composite, the average December through March ECMWF analyses for the winters 1979, 1985, 1986, 1987, and 1988 were used. Only three of these winters have negative index values, resulting in a composited index value of -0.6 ± 0.8. The high NAO composited index is 3.5 ± 0.9, determined from the average of the 1983, 1989, 1990, 1992 and 1993 winter indices.

The transient eddy statistics come from ECMWF analyses filtered to retain fluctuations between 2 and 8 days using the bandpass filter described by Trenberth (1991). The term ‘storm tracks’, therefore, refers to regions of maximum variance arising from disturbances with periods less than about one week. The distribution of the 2 to 8 day-filtered standard deviations of 300 mb geopotential height for

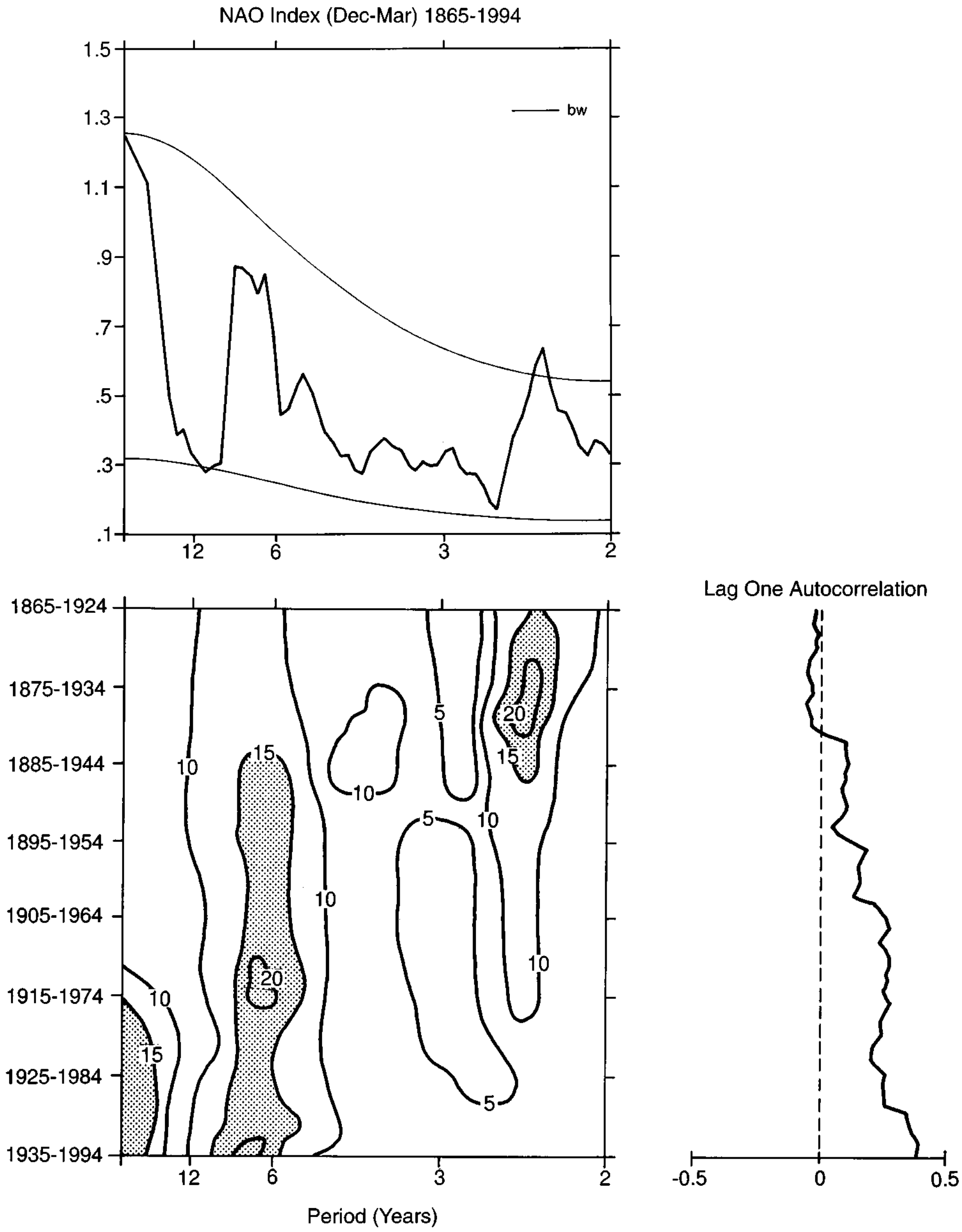

Figure 7. Power spectra of the winter (December–March) NAO index for 1865–1994 (top) and running 60-year intervals (bottom). Variances greater than 0.15 mb^2 frequency^{-1} are stippled in the lower panel. Also shown is the lag one autocorrelation coefficient for each 60 year interval.

the December–March winter season averaged over the years 1979–1994 reveals the well-known variance maxima over the North Pacific and Atlantic oceans (Figure 10). The difference high minus low NAO index values shows a clear shift in

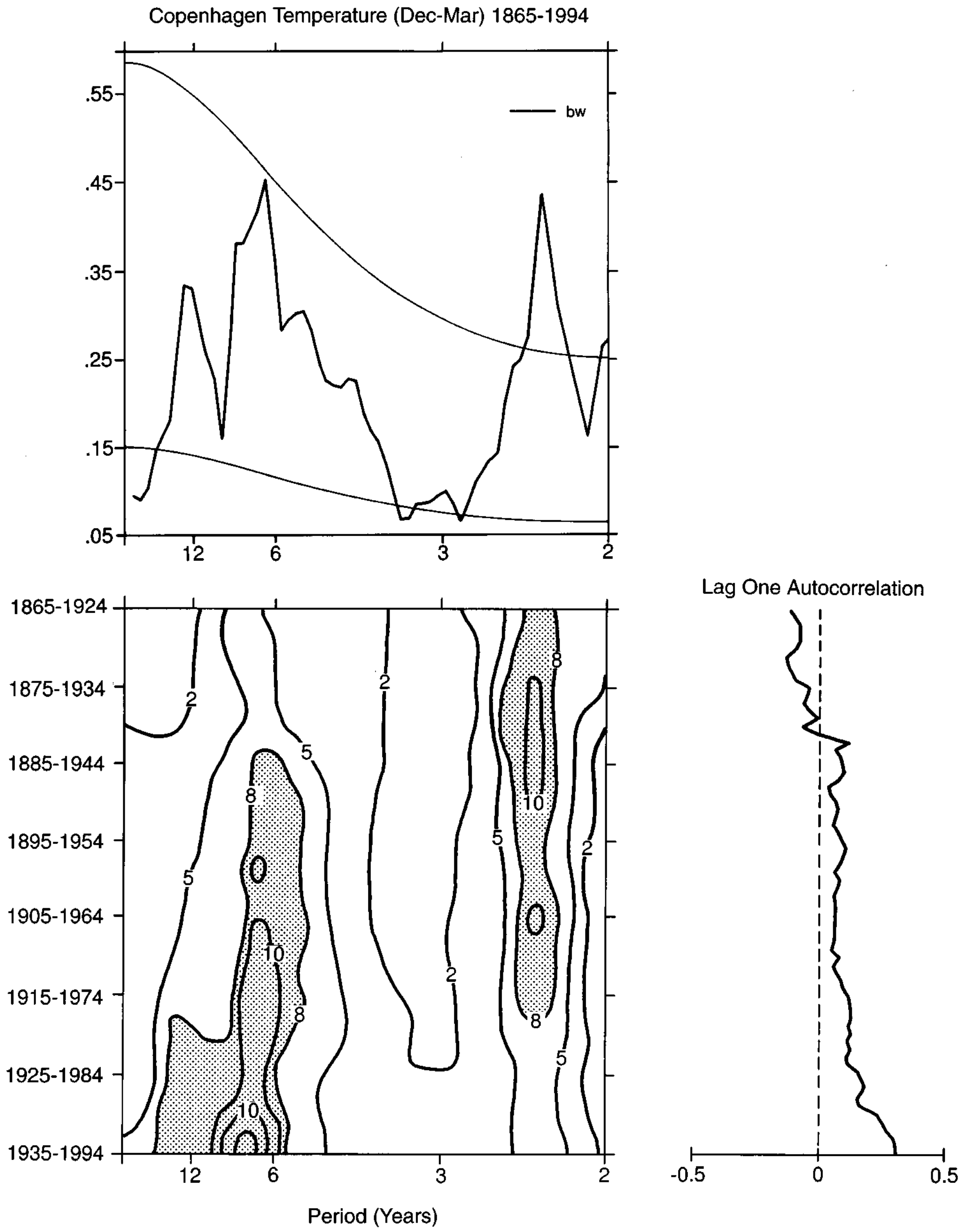

Figure 8. As in Figure 7 but for Copenhagen winter temperature anomalies. Variances greater than 0.08 $^{\circ}C^2$ frequency^{-1} are shaded in the lower panel.

storm track activity with statistically significant enhanced variance over the North Atlantic and northern Europe and reduced activity over the subtropical Atlantic.

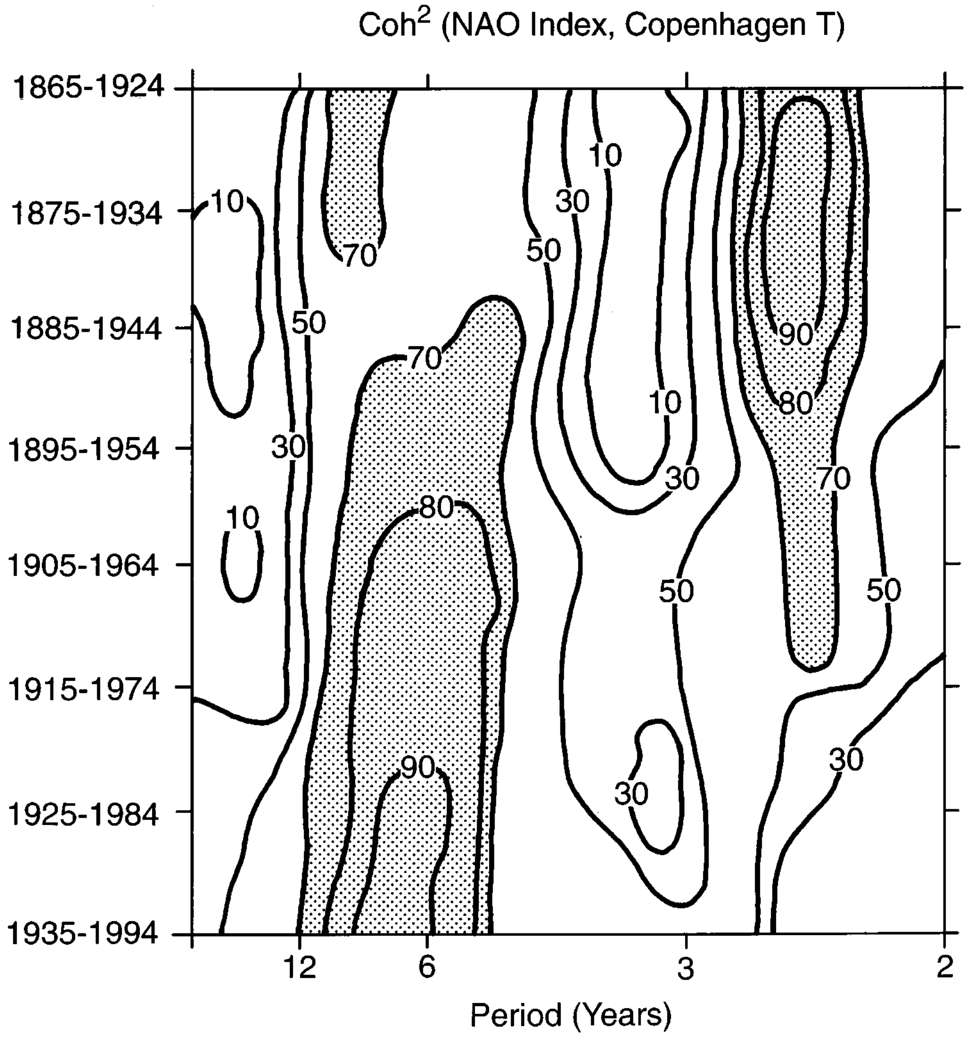

Figure 9. Coherence squared ($\times$ 100) between the winter (December–March) NAO index and Copenhagen temperature anomalies over running 60 year intervals. Values greater than 70 are shaded.

The systematic change in the storm tracks allows the possibility that anomalous transient heat and vorticity (or momentum) fluxes may help maintain the anomalous mean circulation. The total direct eddy forcing of the mean streamfunction (ψ) is best represented by the term

$$\frac{\partial\overline{\psi}}{\partial t} = -\nabla^{-2}(\nabla \cdot \overline{\mathbf{v}'_{\psi}\zeta'}) \tag{2}$$

where ζ is the vorticity, $\mathbf{v}_{\psi}$ is the rotational component of the wind, overbars represent the mean flow, and primes represent the bandpass time-filtered transient flow (Hurrell, 1995b). At 300 mb (Figure 11) the dominant features for the high

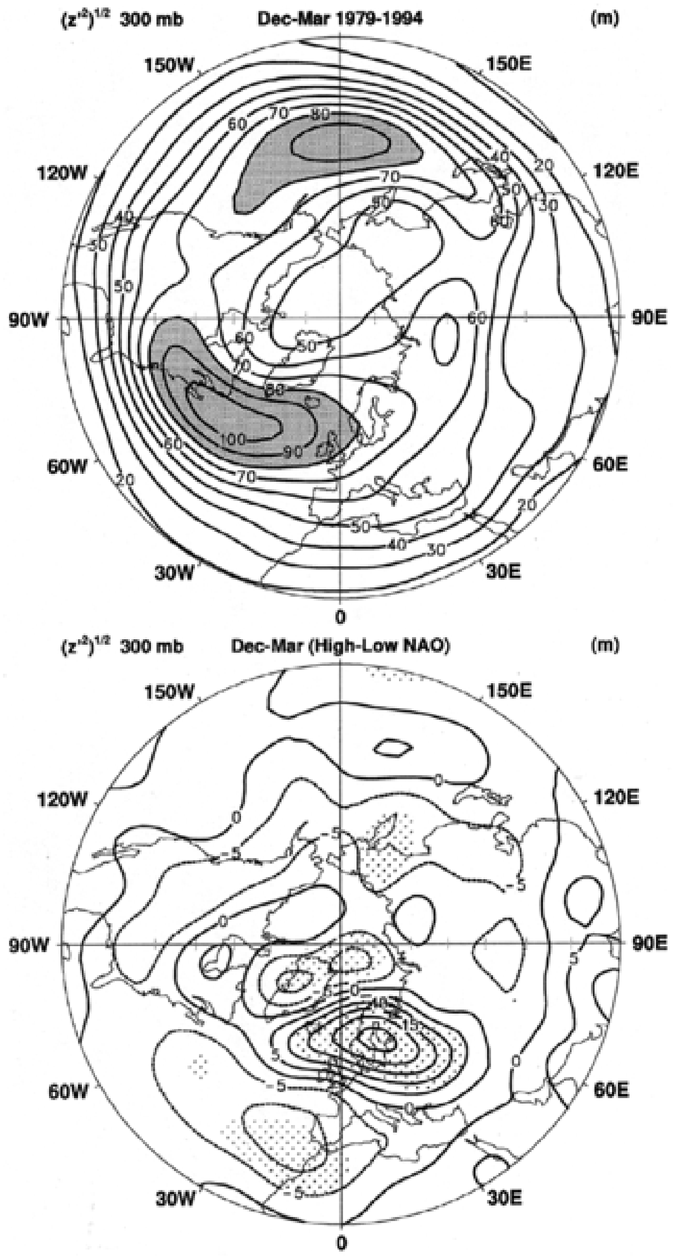

Figure 10. Mean storm tracks for 1979–1994 winters (December–March) and the anomalies, high–low NAO index winters, as revealed by the 300 mb root mean square transient geopotential height $\overline{(z'^2)}^{1/2}$ (m) bandpassed to include 2 to 8 day period fluctuations. Values greater than 80 m are indicated by dark shading in the top panel. Values significantly different from zero at the 5% level using a *t* test are stippled on the lower panel. Results have been smoothed to T21 resolution.

minus low NAO index composite are a large cyclonic circulation forcing over the far north Atlantic and strong anticyclonic forcing over the middle and subtropical latitudes of the Atlantic and Europe, which coincides with the anomalous mean circulation at this level (see Figure 9 of Hurrell, 1995b). This shows that the transient eddies are systematically reinforcing and helping to maintain the upper tropospheric rotational flow in its anomalous form for a large positive NAO index. Similar results from linear model analyses of the role of transient eddy vorticity fluxes to the maintenance of anomalous extratropical wavetrains have been noted in many previous studies, mostly dealing with the Pacific (e.g., Held et al., 1989; Hoerling and Ting, 1994). A similar pattern is noted in the lower troposphere (Figure 11) but the tendencies are much smaller and the effects of the anomalous transient heat fluxes are more important.

The effects of the anomalous transient eddy heat fluxes on the mean circulation can be examined through the thermodynamic equation as the tendency in the mean temperature field

$$\frac{\partial \overline{T}}{\partial t} = -\nabla \cdot \overline{\mathbf{v}'T'} \tag{3}$$

Figure 12 shows the 700 mb perturbation mean temperature for the composited high minus low NAO index differences and the heat flux divergence from Equation (3). It reveals a striking negative correlation over the Atlantic and Europe which implies that the high frequency transient eddies are acting to destroy the mean temperature perturbation in a diffusive manner over these regions. This illustrates that it is the advection by the mean flow that is offsetting the eddy forcing and maintaining the temperature perturbation. This finding is compatible with the view that the transients are baroclinic eddies influenced by the anomalous temperature gradients on which they feed to produce downgradient transports (e.g., van Loon, 1979; van Loon and Williams, 1980). Consequently, the role of the transient eddy forcing is mixed. In the lower troposphere the high frequency transients act to interfere with the anomalous mean flow while in the upper troposphere, the vorticity fluxes are dominant and act to reinforce the anomalous circulation. It is clear that the change in storm tracks plays a significant role in shaping the anomalous mean pattern.

e. Relationships to Precipitation

The changes in the mean and eddy components of the flow affect the transport and convergence of moisture and, therefore, can be directly tied to changes in regional precipitation. Since the early 1980s conditions have been anomalously dry over southern Europe and the Mediterranean and wetter-than-normal over northern Europe and parts of Scandinavia (Figure 13). Over the Alps, for instance, snow depth and duration over the past several winters have been among the lowest recorded this century, causing economic hardships on those industries dependent on winter snowfall (Beniston and Rebetez, 1996).

Hurrell (1995a) used composited ECWMF analyses and showed that, during times of a high NAO index, the axis of maximum moisture transport shifts to a more

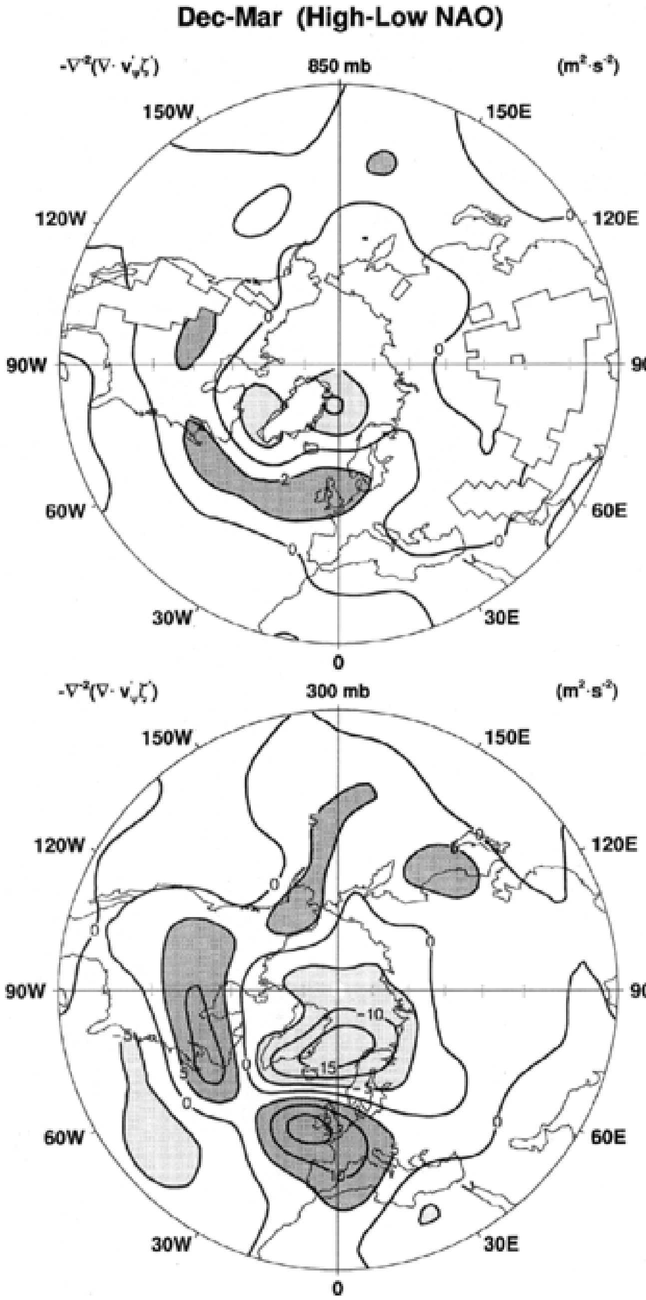

Figure 11. The forcing of the mean streamfunction, high–low NAO winters, associated with the convergence of the vorticity flux by the transient rotational flow at 850 mb and 300 mb. The contour increment in the top panel is 2 $m^2 s^{-2}$ and 5 $m^2 s^{-2}$ in the lower panel. Cyclonic (anticyclonic) tendencies are indicated by light (dark) shading. Results have been smoothed to T21 resolution.

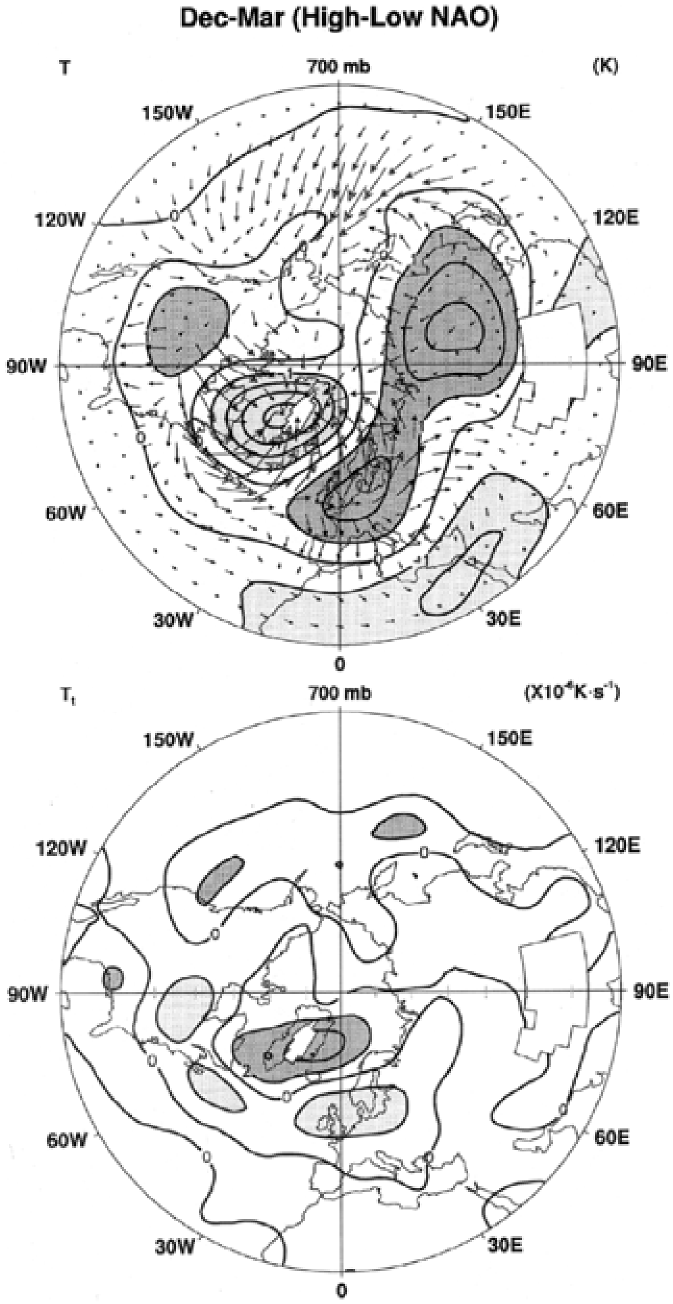

Figure 12. Top: Mean temperature anomalies (K), high–low NAO winters, at 700 mb, along with the anomalous heat flux by the 2–8 day transient eddies. The longest vector represents a heat flux of 5 K m s^{-1}. Temperature anomalies > 1 °C are indicated by dark shading, and those < -1 °C are indicated by light shading. Bottom: Forcing of the anomalous mean temperature by the transient eddy heat flux convergence at 700 mb. Values $> 2 \times 10^{-6}$ K s^{-1} are given by dark shading, and values $< -2 \times 10^{-6}$ K s^{-1} are indicated by light shading.

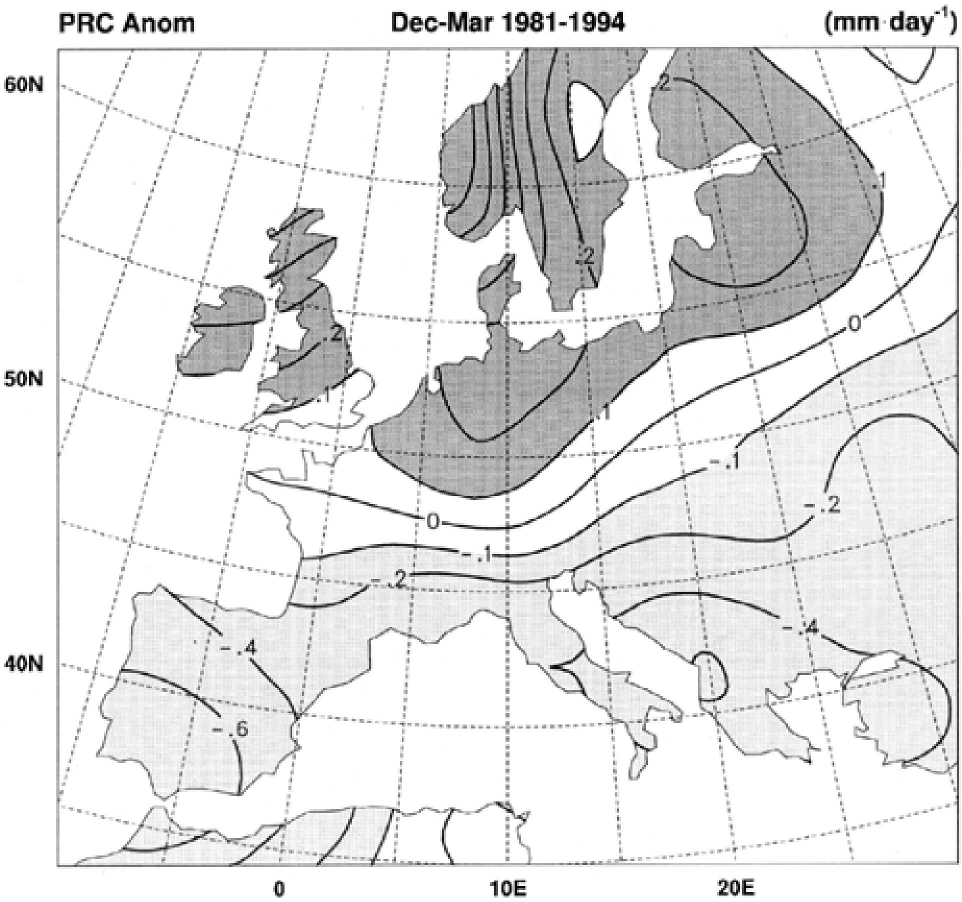

Figure 13. Fourteen winter (1981–1994) average precipitation anomalies expressed as departures from the 1951–1980 mean. The contour increment is 0.2 mm day^{-1}, except the ± 0.1 mm day^{-1} contours are included. Anomalies > 0.1 mm day^{-1} are indicated by dark shading, and those < -0.1 mm day^{-1} are indicated by light shading. The dataset was kindly provided by Dr. Mike Hulme and is an extension of the dataset described by Eischeid et al. (1991).

southwest-to-northeast orientation across the Atlantic and extends much farther to the north and east onto northern Europe and Scandinavia (see his Figure 4). A significant reduction of the total atmospheric moisture transport occurs over parts of southern Europe, the Mediterranean and North Africa. Plots of evaporation minus precipitation ($E - P$), computed as a residual of the atmospheric moisture budget, show that E exceeds P over much of Greenland during high NAO index winters, a result that is consistent with other observational and modeling studies that indicate a declining precipitation rate over much of the Greenland Ice Sheet over the past two decades (Bromwich et al., 1993). Drier conditions during high NAO index winters were also implied by the ECMWF data over much of central and southern

Europe and the Mediterranean, while enhanced moisture flux convergence was indicated from Iceland through Scandinavia.

The changes in precipitation associated with a one standard deviation change in the NAO index support the $E - P$ patterns of Hurrell (1995a) over Europe (Figure 14), and the similarity to the decadal precipitation anomalies in Figure 13 shows that the recent conditions over Europe can be directly linked to the behavior of the NAO. Long-term station data lend further support to the pattern in Figure 14. Listed in Table II are 39 stations that contain December–March records of precipitation for at least 40 winters. Shown are the correlation coefficients with the NAO index in Figure 5 and the number of winters that were included in the correlations. Many of the correlations are highly significant, and those of greatest magnitude generally occur where the departures in Figure 14 are largest. Also indicated in Table II is the mean winter precipitation rate for each station, computed over the total number of years, as well as the difference in precipitation rates for winters in which the NAO index exceeded 1.0 minus winters in which the index was below –1.0. The agreement with Figures 13 and 14 is quite good, and it provides further evidence that the recent precipitation anomalies are related to the persistent positive phase of the NAO since about 1980.

4. Conclusions

Variations in local surface variables on interannual and longer time scales are not uniform, but occur in distinctive large-scale patterns. The spatially coherent patterns arise because the surface variations are associated with changes in the quasi-stationary planetary waves among other factors (see Trenberth, 1993, for a review). An example is the surface temperature and SLP anomalies shown in Figure 1. Studies of long-term trends of temperature or precipitation in local climate records, such as those from high elevation sites, should also examine parallel changes in the circulation in order to fully synthesize the available information about regional or global climate change. The intent of our paper at HIGHEST-95 was to show that recent decadal changes in temperature and precipitation are related to pronounced changes over the past twenty years in the wintertime atmospheric circulation over the ocean basins of the NH.

The pattern of temperature change over the past two decades has been one of warming over NH continents and cooling over the oceans (Figure 1a). Elements of this pattern resemble the greenhouse warming fingerprint predicted by some general circulation models (IPCC, 1996), and some have suggested that the recent global warming may be related to increasing tropical ocean temperatures that have led to an enhancement of the tropical hydrological cycle (e.g., Graham, 1995). Whether the observed changes are in response to greenhouse gas forcing, or whether the changes are part of a natural decadal time scale variation, is difficult to assess and is a matter of much debate.

Table II

Stations (latitude, longitude) that contain records of December–March precipitation for at least 40 winters. The correlation coefficients r(NAO, P) with the NAO index (Figure 5) and the total number of winters (n) that were included in the correlations are indicated. Also indicated are the mean precipitation rate $\overline{P}$ over the total number of winters (n), and the difference in precipitation rate between winters with a NAO index $>$ 1.0 and those with an index $<$–1.0. One asterisk indicates statistical significance at the 5% level and two indicate significance at the 1% level

Station	r(NAO, P)	n	$\overline{P}$(mm day^{-1})	ΔP(mm day^{-1})
Bergen (60.4° N, 5.3° E)	0.77**	72	5.8	3.6**
Stornoway (58.2° N, 6.3° W)	0.75**	63	3.5	1.4**
Tiree (56.6° N, 6.9° W)	0.67**	63	3.4	1.2**
Stavanger (58.9° N, 5.6° E)	0.66**	43	2.7	1.4**
Thorshavn (62.0° N, 6.8° W)	0.53**	116	4.8	1.1**
Lerwick (60.1° N, 1.2° W)	0.49**	63	3.6	0.8**
Reykjavik (64.1° N, 21.9° W)	0.48**	73	2.7	0.9**
Akureyri (65.7° N, 18.1° W)	0.43**	63	1.6	0.4*
Stykkisholmur (65.1° N, 22.7° W)	0.40**	117	2.2	0.7**
Haparanda (65.8° N, 24.2° E)	0.37**	130	1.2	0.4**
Karesuando (68.5° N, 22.5° E)	0.21*	115	0.6	0.1
Oslo (60.2° N, 11.1° E)	0.21*	125	1.3	0.2
Helsinki (60.3° N, 25.0° E)	0.18*	130	1.5	0.1
Edinburgh (56.0° N, 3.4° W)	0.14	130	1.7	0.2
Stockholm (59.4° N, 18.0° E)	0.14	126	1.1	0.1
Copenhagen (55.7° N, 12.6° E)	0.14	129	1.3	0.1
Valentia (51.9° N, 10.3° W)	0.09	123	4.5	0.1
De Bilt (52.1° N, 5.2° E)	0.08	130	1.9	0.1
Belfast (54.7° N, 6.2° W)	0.01	63	2.3	0.0
London (51.2° N, 0.2° W)	–0.02	129	1.7	–0.0
Angmagssalik (65.6° N, 37.6° W)	–0.02	90	2.6	–0.0
Athens (38.0° N, 23.7° E)	–0.11	98	1.7	–0.1
Egedesminde (68.7° N, 52.8° W)	–0.13	38	0.6	–0.1
Paris (49.0° N, 2.5° E)	–0.19*	119	1.5	–0.3**
Frankfurt (50.1° N, 8.7° W)	–0.19*	130	1.4	–0.2*
Godthåb (64.2° N, 51.8° W)	–0.20*	100	1.1	–0.4
Jakobshavn (69.2° N, 51.1° W)	–0.21*	95	0.4	–0.1
Ivigtut (61.2° N, 48.2° W)	–0.31*	89	2.8	–0.9*
Marseille (43.5° N, 5.2° E)	–0.32**	120	1.5	–0.5*
Milan (45.4° N, 9.3° E)	–0.35**	130	2.2	–0.8**
Istanbul (41.0° N, 29.1° E)	–0.36**	64	2.7	–0.7**
Lyon (45.7° N, 5.0° E)	–0.37**	129	1.6	–0.5**
Rome (41.8° N, 12.2° E)	–0.37**	119	2.6	–0.8**
Ajaccio (41.9° N, 8.8° E)	–0.48**	42	2.2	–1.1**
Ponta Delgado (37.8° N, 25.7° W)	–0.49**	98	3.2	–1.1**
Belgrade (44.8° N, 20.5° E)	–0.50**	94	1.4	–0.6**
Casablanca (33.6° N, 7.7° W)	–0.61**	82	2.0	–1.1**
Lisbon (38.8° N, 9.1° W)	–0.64**	130	3.0	–1.8**
Madrid (40.4° N, 3.7° W)	–0.69**	129	1.3	–1.0**

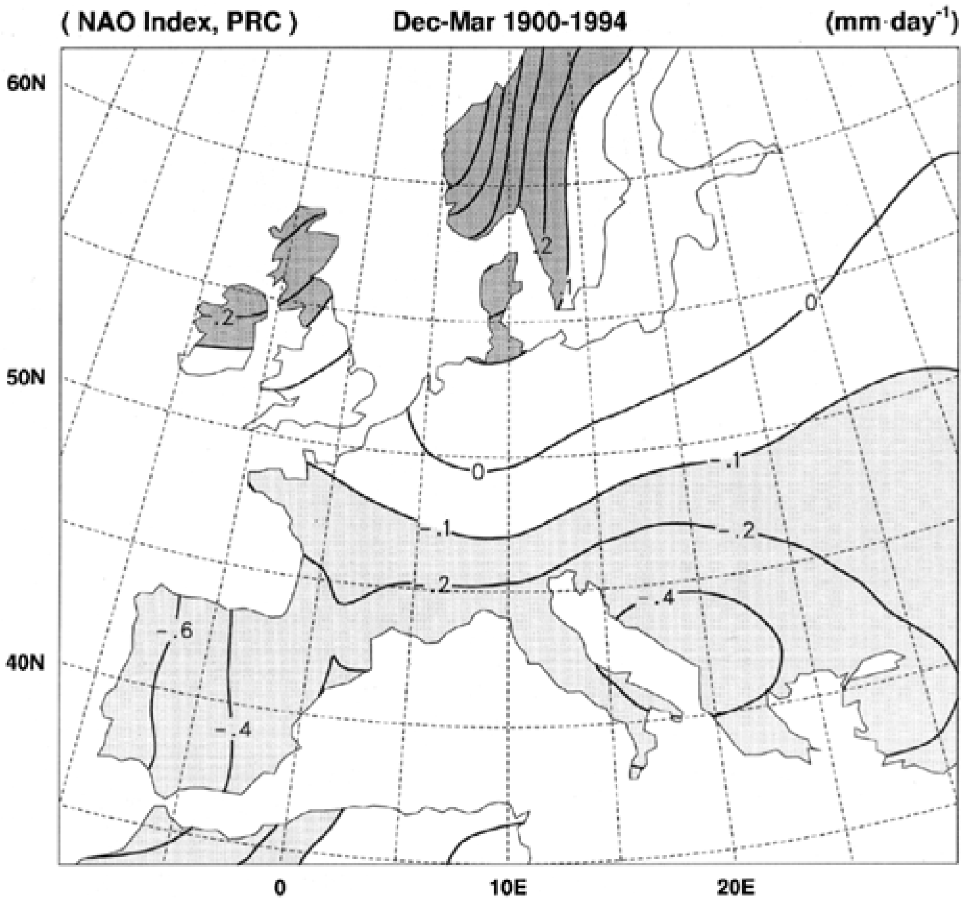

Figure 14. Changes in precipitation corresponding to a unit deviation of the NAO index computed over the winters (December–March) from 1900 through 1994. The contour increment and shading are as in Figure 13.

Palmer (1993) has emphasized that nonlinear dynamics may well change the frequency distribution of weather regimes in the atmosphere, rather than changing the regimes, so that counter-intuitive changes can occur. Trenberth and Hoar (1996) have argued that the tendency for more frequent El Niño events and fewer La Niña events since the late 1970s, which is linked to the decadal changes over the North Pacific, may be an example of Palmer's arguments and thus the changes may be a manifestation of global warming and related climate change associated with increases in greenhouse gases in the atmosphere. On the other hand, abrupt decadal changes in the climate of the Atlantic have been revealed in analyses of ice core data from Greenland (e.g., Alley et al., 1993) that may be related to natural fluctuations in the NAO (Barlow et al., 1993) resulting from internal atmospheric dynamical processes (e.g, Barnett, 1985; Wallace et al., 1995). Regardless of cause, the changes in circulation over the past two decades have resulted in a particular surface temperature anomaly pattern that has amplified the hemispheric-averaged warming over the NH because of its interaction with land and ocean. At the surface, temperature anomalies over land are accentuated relative to the oceans due to the larger heat capacity of water and the depth of the layer linked to the surface (Hurrell and Trenberth, 1996). The result is that the hemispheric mean surface air temperature is largely determined by the temperature of the continents.

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