Seasonality of the North Atlantic Oscillation

Diane H. Portis

Cooperative Institute for Mesoscale Meteorological Studies, University of Oklahoma, Norman, Oklahoma, and Department of Atmospheric Sciences, University of Illinois at Urbana–Champaign, Urbana, Illinois

John E. Walsh

Department of Atmospheric Sciences, University of Illinois at Urbana–Champaign, Urbana, Illinois

Mostafa El Hamly

Cooperative Institute for Mesoscale Meteorological Studies, University of Oklahoma, Norman, Oklahoma, and Moroccan Direction de la Météorologie Nationale, Casablanca, Morocco

Peter J. Lamb

Cooperative Institute for Mesoscale Meteorological Studies, and School of Meteorology, University of Oklahoma, Norman, Oklahoma

(Manuscript received 11 April 2000, in final form 2 October 2000)

ABSTRACT

Monthly sea level pressure (SLP) data from the National Centers for Environmental Prediction reanalysis for 1948–99 are used to develop a seasonally and geographically varying ‘‘mobile'' index of the North Atlantic oscillation (NAOm). NAOm is defined as the difference between normalized SLP anomalies at the locations of maximum negative correlation between the subtropical and subpolar North Atlantic SLP. The subtropical nodal point migrates westward and slightly northward into the central North Atlantic from winter to summer. The NAOm index is robust across datasets, and correlates more highly than EOF coefficients with historical measures of westerly wind intensity across North Atlantic midlatitudes. As measured by this ‘‘mobile index,'' the NAO's nodes maintain their correlation from winter to summer to a greater degree than traditional NAO indices based on fixed stations in the eastern North Atlantic (Azores, Lisbon, Iceland). When the NAOm index is extended back to 1873, its annual values during the late 1800s are strongly negative due to negative contributions from all seasons, amplifying fluctuations present in traditional winter-only indices. In contrast, after the mid-1950s, the values for different seasons sufficiently offset each other to make the annually averaged excursions of NAOm smaller than those of winter-only indices. Global teleconnection fields show that the wider influence of the NAO—particularly in the western North Atlantic, eastern North America, and Arctic—is more apparent during spring–summer–autumn when the NAOm is used to characterize the NAO. Thus, the mobile index should be useful in NAO investigations that involve seasonality.

1. Introduction

The North Atlantic oscillation (NAO) has long been recognized as a major large-scale mode of the atmospheric circulation over the extratropical ocean between North America and Europe (e.g., Walker and Bliss 1932; van Loon and Rogers 1978; Wallace and Gutzler 1981; Lamb and Peppler 1987; Hurrell 1995). Most investigations of the NAO have focused on interannual-to-decadal timescales, particularly in association with short-term climate variations over Europe and North America (Dickson and Namias 1976; Cheng et al. 1995; Hurrell and van Loon 1997) and with extratropical ocean variations (Deser and Blackmon 1993; Kushnir 1994; Dickson 1997). Ward et al. (1999) have investigated the intraseasonal evolution of the NAO for seasonal precipitation prediction in northwest Africa. Recently, the NAO has been set in a broader spatial context of hemispheric variations, both at the surface and aloft (Thompson and Wallace 1998; Thompson et al. 2000; Wallace 2000). As shown in the latter studies, an annular mode (the ‘‘Arctic oscillation'') encompasses many of the features of the NAO, at least during the winter portion of the year.

Corresponding author address: Diane Portis, Department of Atmospheric Sciences, University of Illinois at Urbana–Champaign, 105 South Gregory Avenue, Urbana, IL 61801-3070.
E-mail: portis@atmos.uiuc.edu

The studies cited above have generally focused on the winter season, when the NAO is most clearly defined in terms of its traditional measures. The most widely used measure is the difference of normalized sea level pressure (SLP) anomaly between Iceland and the subtropical eastern North Atlantic. Specific locations used for index definition include Akureyri, Iceland, and Ponta Delgada, Azores (Rogers 1984) and Stykkisholmur, Iceland, and Lisbon, Portugal (Hurrell 1995). The subtraction of the normalized subpolar SLP anomaly from the normalized subtropical SLP anomaly results in a standardized index that is positive when there is an enhanced pressure difference (pressure gradient) between the two centers. A strengthening (weakening) of the westerlies is a consequence of a positive (negative) NAO index.

Several studies have addressed North Atlantic atmospheric circulation variability in terms of empirical orthogonal functions (EOFs) or rotated principal components (Wallace and Gutzler 1981; Barnston and Livezey 1987; Glowienka-Hense 1990; Davis et al. 1997). Such approaches provide insight into the seasonality of dominant modes. However, the orthogonality constraint and use of a hemispheric domain often prevent a clear focus on a particular regional mode such as the NAO—because each pattern is constrained to be orthogonal to all the others. The above studies nevertheless suggest that primary features of NAO-like modes shift seasonally. For example, the subtropical node of the North Atlantic dipole in a leading principal component is generally located farther west during summer than winter (Barnston and Livezey 1987; Glowienka-Hense 1990).

Given the major role of the NAO in determining weather and climate variability over the Atlantic and European sectors, a clear delineation of the NAO's seasonality is needed. In this paper, we evaluate the NAO's seasonality in an NAO-specific framework that avoids complications introduced by dynamical or statistical ties to other modes of variability. This approach to the NAO is intended to 1) optimize the NAO index for applications in monitoring and seasonal prediction, and 2) provide a NAO-specific index for studies of the seasonality of broader modes of variability encompassing the NAO, for example, the Arctic oscillation.

2. Methodology and data

The framework we chose for a seasonally dependent NAO index was the identification of the NAO nodal locations that anchor coherent areas of maximum negative correlation of SLP in each calendar month. Two broad boxes encompassing potential nodal locations were selected. Both boxes spanned the same longitudinal range (70°W–0°), with the latitudinal range for the northern (southern) node being 55°–80°N (20°–45°N). We do not extend the domain eastward from 0° because, although some EOF analyses contain loadings farther east in NAO-like patterns (e.g., Barnston and Livezey 1987), the mean positions of the North Atlantic subpolar low and subtropical high are west of 26°W in all calendar months (Mächel et al. 1998, their Fig. 2). For each calendar month, the SLP value at each grid point within the northern box was correlated with each gridded SLP value in the southern box. The two most negatively correlated north–south points were selected as the NAO nodes for that calendar month. If more than one pair of nodes had similarly high negative correlations (i.e., the correlations differed by less than 0.05), the final selection criterion was proximity to the long-term monthly mean locations of the subpolar low and the subtropical high.

The principal gridded dataset used to determine the location of the NAO nodes was the monthly averaged SLP fields of the National Centers for Environmental Prediction (NCEP) reanalysis (Kalnay et al. 1996) provided by the National Oceanic and Atmospheric Administration–Cooperative Institute for Research in Environmental Science (NOAA–CIRES) Climate Diagnostics Center, Boulder, Colorado (http://www.cdc.noaa.gov). This monthly dataset has a horizontal resolution of 2.5° lat $\times$ 2.5° long. The primary set of computations in this study is based on the 52-yr period (1948–99) of the NCEP reanalysis. The advantages of a reanalysis dataset include internal consistency among model variables in time and space due to the "frozen" model physics and computational schemes, and the incorporation of a large set of available atmospheric data from diverse sources. This latter advantage is *potentially* important over the North Atlantic, where the regular synoptic observing network is sparse. However, an inherent risk in using reanalysis output is that the nature and quantity of input data streams may change over time, thereby introducing temporal inhomogeneities into the derived product. In the case of the NCEP reanalysis, it is known that there were fewer input observations during the first 10 years (1948–57), which were not included in the original NCEP reanalysis (Kalnay et al. 1996). In section 3, we address this issue by examining the means and variances of the NCEP reanalysis SLP at the NAO nodal points for the periods before and after 1 January 1958. We also use the National Center for Atmospheric Research (NCAR) (Trenberth and Paolino 1980) and University of East Anglia (see below) SLP data in section 3 to further confirm the stability of our results.

A disadvantage of the NCEP reanalysis data for low-frequency analysis is that the data only extends back to 1948. We used the Northern Hemispheric gridded SLP dataset of Jones (1987) to obtain monthly time series of SLP variations at the reanalysis-derived nodal locations back to 1873. This dataset, which is available from the Climatic Research Unit of the University of East Anglia, Norwich, United Kingdom (http://www.cru.uea.ac.uk/cru/data/pressure.htm), has a coarser grid resolution (5° lat $\times$ 10° long) than the NCEP reanalysis data. Hence we used the closest available grid points to the reanalysis-derived mobile node locations when we constructed NAOm (1873–1995) from the Jones SLP dataset.

The ''mobile'' NAO index (NAOm) is defined as the difference between the normalized SLP departures at the calendar-month-dependent nodes (southern minus northern). For comparative purposes, the traditional Hurrell (1864–1997) and Rogers (1874–1999) indices, defined above, are also included in this paper. The normalization period used in the calculation of the Rogers index was its period of record, while the Hurrell index is based on a normalization period of 1864–1983 (Hurrell 1995). The station data for the calculation of the Hurrell index are available online (http://www.met.rdg.ac.uk/cag/NAO/index.html). The station data for the calculation of Rogers' index were obtained from Jeffrey Rogers (The Ohio State University) in 1994 and have been updated from Monthly Climatic Data of the World (National Climate Data Center, Asheville, North Carolina). The Rogers and Hurrell indices were developed and used specifically for winter conditions, so they may not capture the key variability during the nonwinter months. In order to emphasize the winter-only (W) focus of these traditional indices, we denote the Hurrell and Rogers indices by W_H and W_R, respectively, in our results.

As stated above, we used the raw data from the Jones dataset for the low-frequency analysis of the mobile NAO index. Using an 1873–1995 normalization period, normalized departures were obtained for that entire period from seasonal and annual means and their standard deviations. The NAOm index constructed from the difference of these normalized departures was filtered with a centered 11-term Lanczos filter with a frequency cutoff of 0.1 cycles yr^{-1} value (Duchon 1979). The Lanczos filter has a sharper response curve than a moving average filter, due to a reduction in Gibbs oscillations. This reduction in Gibbs oscillations is accomplished by multiplying the original weight function by a ''sigma factor.'' For our filter, this sigma factor was raised to a power of 1.0. The five filtered values at either end of the time series were approximated by applying the same 11-term filter to the available data, effectively creating uncentered estimates.

3. Results

a. Annual cycle characteristics of NAO

Figure 1 shows the calendar-month locations of the points that define the NAOm described in section 2, along with the ''traditional'' fixed nodal points of the wintertime indices of Hurrell (W_H) and Rogers (W_R). The mobile subtropical high nodes (computed for 1948–99) are mainly clustered in the center of the Atlantic basin, with the exception of June and July (located over the Gulf Stream) and January (eastern Atlantic). There is a general southeast–northwest migration of the subtropical node, from winter to summer, and a return from summer to winter, with the exception of southwestward excursions in February–March and especially November. For November, a subtropical–subpolar coupled node also was identified in the eastern Atlantic (not shown), but its correlation was smaller than the central Atlantic correlation (Fig. 1) by more than 0.05 (see section 2). The annual march of the subtropical node has an east–west range of approximately 3000 km, thus making the use of a point in Portugal or the Azores to characterize the NAO less than optimal for certain months (e.g., June, July, November). The mobile subpolar and subtropical nodes undergo nearly ''parallel'' month-to-month displacements, so the latitudinal separation between the two nodes generally remains constant.

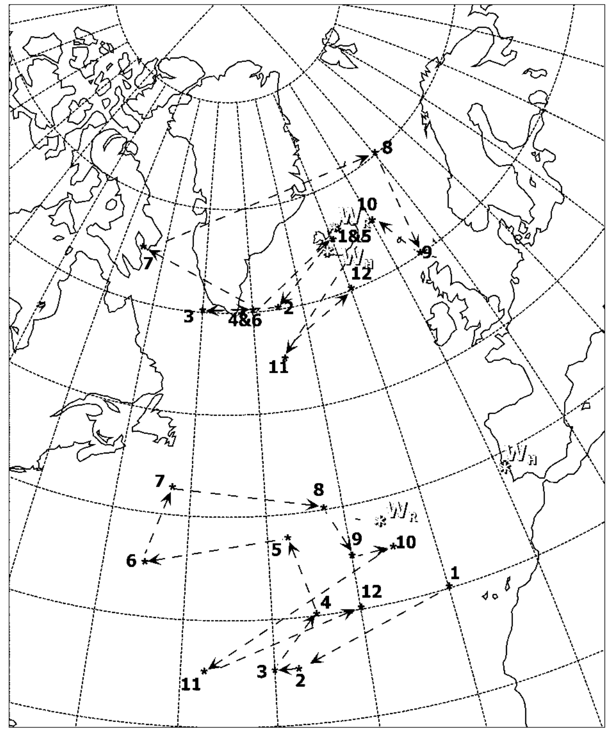


FIG. 1. Locations of nodes defining the three NAO indices considered—Mobile, Rogers, and Hurrell. The monthly nodal locations of the Mobile index (computed for 1948–99) are labeled by month number; the station-based nodes of the Rogers's and Hurrell's indexes are located by ''W_R'' and ''W_H,'' respectively. Arrows indicate the annual march of the mobile NAO index nodes.

The general seasonal migration of the nodal NAOm locations in Fig. 1 shows some qualitative similarity to the migration of the climatological centers of the Icelandic low and Azores high (Hastenrath 1991, p. 137; Mächel et al. 1998) in the sense that the locations are generally farther south in the winter than in the summer. However, there are large differences in some months—for example, the November subtropical node in Fig. 1 is approximately 2000 km farther south and west than the center of the subtropical high in November, and the June–July nodes are farther west than the locations of the corresponding climatological pressure centers. In particular, the July subpolar node in Fig. 1 is much farther west than the Icelandic low's climatological position (63°N, 27°W) in July. These differences reveal that the largest internode NAO anomaly correlations often do not occur in the immediate vicinity of the climatological centers of sea level pressure.

The general westward shift of the subtropical node from winter to summer was also found by Sahsamanoglou (1990). A similar tendency is apparent in the 700-mb principal component patterns of Barnston and Livezey (1987), whose NAO-like modes for January, March, May, June, September, and October have nodal points that nearly coincide with those in Fig. 1. However, for July and August, Barnston and Livezey's subtropical nodes split into two centers of comparable magnitude—one over the Great Lakes of North America and the other in the far eastern North Atlantic near the British Isles. Since these patterns were obtained after an orthogonal rotation of hemispheric fields, the final patterns are constrained statistically with respect to the other rotated patterns, some of which likely include North Atlantic features.

Section 2 acknowledged the possibility of inhomogeneities in the reanalysis data due to varying input streams. In the case of the NCEP reanalysis, it is known that there were fewer input observations during the first 10 yr (1948–57). However, the annual mean pressures at our subpolar and subtropical NAOm nodes are 1006.5 and 1020.7 mb, respectively, for 1958–99, which are almost identical to the corresponding values for the 1948–57 decade of fewer data, (1006.0 and 1020.3 mb). Concerning a more important indicator of data homogeneity is the variability of the daily pressures about the calendar-month means, the situation is equally reassuring—the standard deviations at the subpolar node are 11.9 mb for 1948–57 and 12.4 mb (1958–99); while those at the subtropical node are 4.6 mb (1948–57) and 5.0 mb (1958–99). Although these standard deviations are slightly smaller for the earlier period of fewer input observations, the differences are much smaller than the month-to-month changes of the climatological means and standard deviations. Accordingly, the previous discussion of the seasonal migration of NAOm nodes applies equally well to nodes based on data for 1958–99 only, or on data for the entire 1948–99 reanalysis period.

Even though reanalysis-derived SLP values at NAOm nodal points are thus relatively homogeneous for 1948–99, the robustness of the nodal points with respect to both the source and the time period of the data are also potential issues of concern. Some temporal dependence may be expected on the basis of Mächel et al.'s (1998, their Fig. 3) finding of low-frequency variations in the locations and intensities of the Icelandic low and Azores high since 1881. We therefore repeated the computations leading to Fig. 1 with sea level pressure datasets for 1958 onward from the University of East Anglia (UEA; Jones 1987, updated) and NCAR (Trenberth and Paolino 1980, updated), as well as the NCEP reanalysis already mentioned. In addition, we repeated the UEA computations for two additional 40-yr intervals: (1878–1917 and 1918–57) to span the period studied by Mächel et al. (1998). Except for May, the patterns in Fig. 1 changed little when the UEA, NCAR, and NCEP data for 1958 onward were used in place of the NCEP data from 1948. For May, the subtropical node in the NCAR dataset was displaced 12.5°–17.5° farther west from the corresponding NCEP and UEA locations. However, changes to Fig. 1 were more substantial when analysis of UEA data was performed for the earlier time periods. The March–April location of the coupled node in the central Atlantic was shifted to the west in the earlier time periods (especially 1878–1917), whereas the June–July western Atlantic location was not found in the earlier years. Another difference was the westward shift in the earlier years of the subpolar node in May (especially from 1878 to 1917), August, and November. Comparing 1878–1917 and 1918–57, there were only two months (February, July) when dominant coupled nodes differed substantially. During February, the 1918–57 time period had a very strong coupled node in the eastern Atlantic, unlike for the other time periods and datasets considered. The general conclusion of these sensitivity tests is that the nodal NAOm locations are robust with respect to the SLP data source for a single period of 40–50 yr, but that the nodal locations can change from one such period to another. This conclusion is consistent with the findings of Corti et al. (1999) and Hilmer and Jung (2000).

To further elucidate the annual cycle of the NAO, Fig. 2 shows the annual cycle of the monthly correlations between the SLP at its nodes for the 1948–97 reanalysis period. Figure 2 contains separate correlation curves for the mobile NAOm index described in section 2, for Rogers' Azores-based index (W_R), and for Hurrell's Lisbon-based index (W_H). The mobile index and, to a slightly lesser extent, the Rogers index clearly capture the NAO signal in all seasons, with the NAOm correlation being strongest in every calendar month. Both of these indices (especially W_R) indicate that the nodal correlation weakens somewhat in summer, but that the oscillation is present year-round. The W_H index, on the other hand, is as strong as W_R during the winter (the focus of Hurrell's studies), but it weakens rapidly thereafter to vanish during June and July. The annual mean values of the internode correlations in Fig. 2 are -0.66 for NAOm (M), -0.51 for W_R, and -0.32 for W_H.

b. Validation of two-point NAO index

A fundamental issue for the present study was whether *any* two-point index could usefully represent the NAO. While Fig. 2 indicates that seasonal mobility is potentially important in a two-point representation, the results presented so far do not eliminate the possibility that a two-dimensional pattern framework (e.g., EOFs)

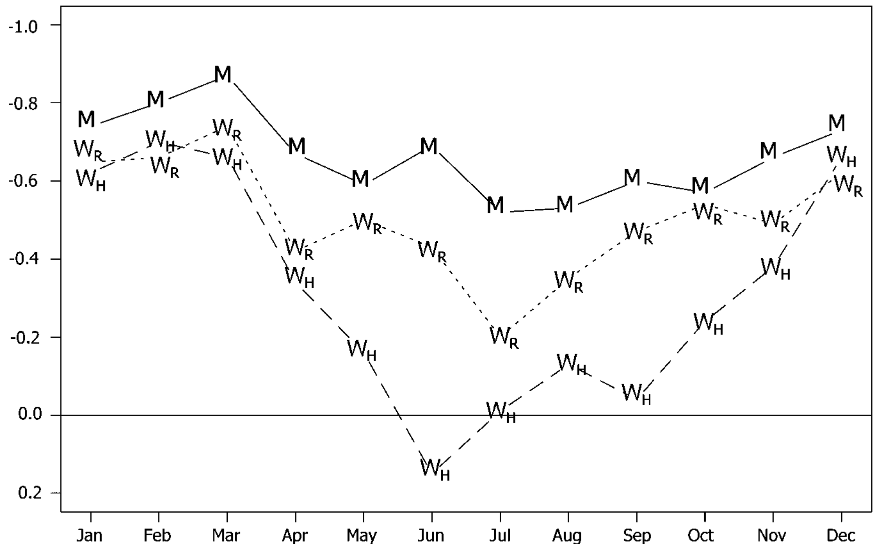


FIG. 2. Monthly internode correlations of the Mobile (M), Rogers (W_R), and Hurrell (W_H) NAO indices for 1948–97. The 99% (95%) significance level is 0.36 (0.28).

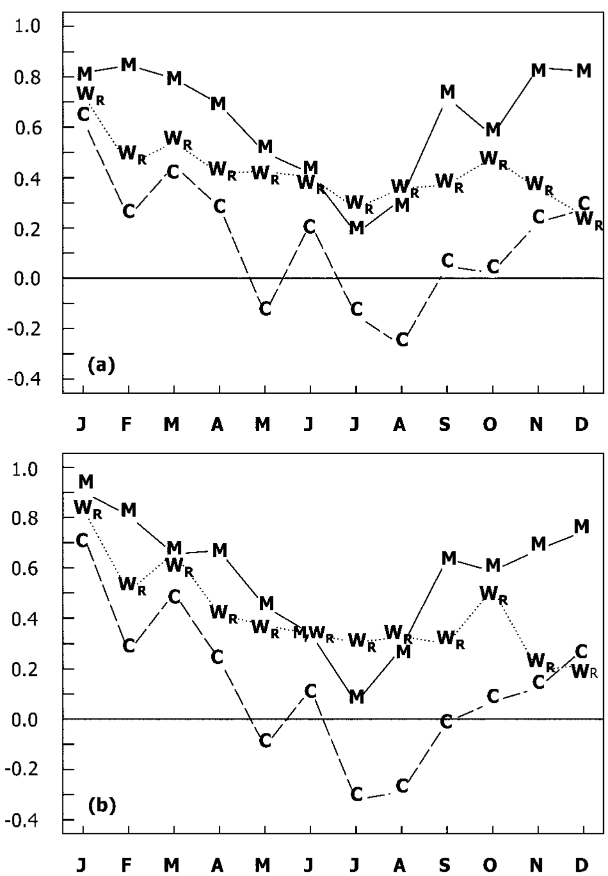
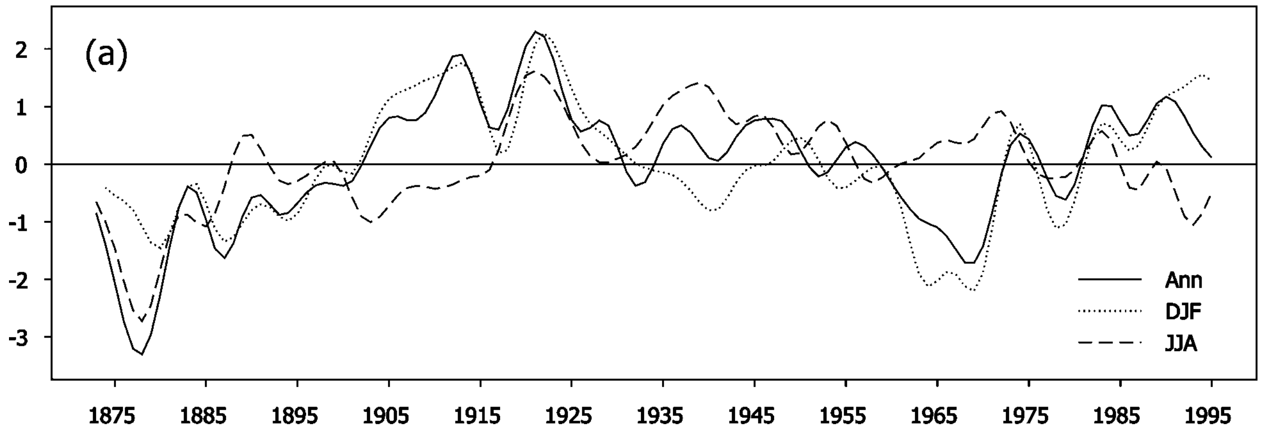


FIG. 3. Monthly correlations of the Mobile (M), Rogers (W_R), and an EOF-based index from NOAA's Climate Prediction Center (C) with westerly indices of (a) Barry and Perry and (b) Lamb for the period 1958–99. The 99% (95%) significance level is 0.39 (0.30).

may be required to capture the essence of NAO variations. A key practical association between the NAO and weather/climate is realized via the intensity of surface westerly airflow across the North Atlantic and into Europe. The variations of this flow have long been a focus of European weather forecasters and climatologists. We have tested the adequacy of our two-point mobile index (NAOm) by evaluating its ability to capture the North Atlantic surface westerly (zonal) airflow. Specifically, we have correlated the mobile NAOm index with (i) Barry and Perry's (1973) index of westerlies over the 35°–55°N, 70°W–10°E region and (ii) a similar westerly index used by Lamb (1972). We also correlated these two westerly indices with Rogers' winter NAO index (W_R) and the EOF-based NAO index maintained by NOAA's Climate Prediction Center (CPC; available online at ftp://ftp.ncep.noaa.gov/pub/cpc/wd52dg/data/indices/tele_index.nh). For the CPC index, the individual monthly EOF coefficients are based on Barnston and Livezey's (1987) 12 calendar-month sets of EOFs, in which the leading EOF is an NAO mode for 7 of the 12 months. The first EOFs of February, April, September, October, and November are not NAO modes; in each of these cases, we used the first subsequent EOF that clearly contained an NAO dipole.

Figure 3a shows that the Barry–Perry (BP) westerly index correlates more highly with NAOm than with W_R and CPC during winter; correlations with NAOm and W_R are similar for summer. The BP–NAOm correlations range from 0.79–0.86 during November–March to 0.18–0.50 during May–August. The lower summer correlations are likely related to the tendency for zonal winds to weaken and migrate northward during that season (Peixoto and Oort 1992, p. 201), sometimes to north of 55°N. In all months, however, the BP–NAOm correla-



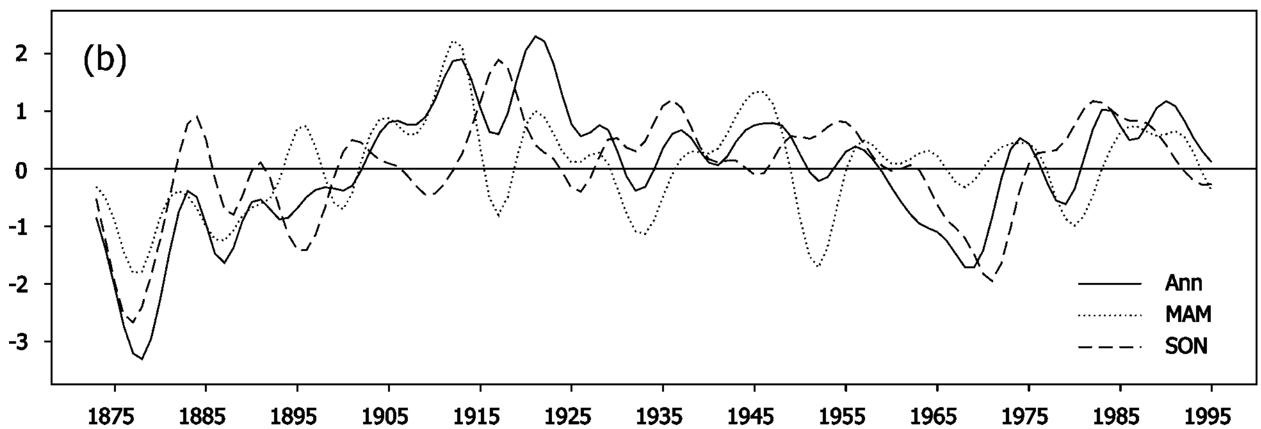


FIG. 4. Low-frequency time series of the Mobile NAO index for contrasting seasons of (a) winter (DJF)–summer (JJA) and (b) spring (MAM)–autumn (SON). Superimposed is the low-frequency annual time series.

tions are substantially stronger than the BP–CPC correlations. Counterpart correlations obtained using the Lamb index are very similar (Fig. 3b). We conclude that, at least for representations of the North Atlantic surface westerlies over the above domain, the EOF representation does not add to the utility of the two-point indices. This conclusion does not necessarily extend to associations with midlatitude westerlies over adjacent longitudinal sectors, that is, those west of 70°W and east of 10°E.

c. Temporal variation of NAO

Low-frequency time series of the NAOm index developed from the UEA–Jones (1987) dataset for opposite seasons (winter/summer and spring/autumn) are shown in Fig. 4, along with a low-frequency annual time series. The annual time series has a century-scale variation that began with a strong negative epoch in the 1870–80s, after which it was positive for an extended period (1900–50s) before transitioning into a negative epoch in the 1960s. The latter was followed by an upward trend in the 1980s. The strongest of the two negative epochs in the annual mean NAOm occurred between 1875 and 1885, due primarily to summer and autumn, while winter had the weakest negative low-frequency signal in those years. The annual mean values remained negative until 1900, although not as strongly as in the 1875–85 period. This negative epoch of the late 1800s is less evident in previously published NAO time series based on winter data (e.g., Hurrell 1995), although it is evident in Sahsamanoglou's (1990, his Fig. 8) time series of annual and July pressure gradients between the Azores high and Icelandic low. The other major negative epoch occurred in the 1960s, where the modulation of the annual values was primarily due to winter and autumn. Unlike the summer–autumn-modulated negative epoch of the late 1800s, this autumn–winter-modulated negative event of the 1960s was apparent in previously published winter-based NAO time series (e.g., Hurrell 1995; Rodwell et al. 1999). The extended period of positive annual values (1900–50s), that is bounded by these two negative epochs, had its largest values before 1930. From 1900 to 1914, winter and spring contributed to the strong positive annual values, while summer and to a lesser extent autumn decreased their strength. The upward trend of the annual mean NAOm following the negative event of the 1960s culminated in the well-publicized recent positive excursion of the 1980s and 1990s (Hurrell 1995; Rodwell et al. 1999). This episode of positive annual values is mainly attributable to positive winter and spring contributions, while the negative summer values since about 1990 weakened the overall positive annual values. A common thread of these low-frequency epochs is a coupling between the seasons that dominates the annual cycle (e.g., summer–autumn for the negative event of 1875–85; winter–spring for the positive event of 1900–14; autumn–winter for the negative event of the 1960s).

It is interesting to note in Fig. 4 that the NAOm values for the contrasting seasons of winter–summer and spring–autumn are often negatively correlated. This is particularly true of winter–summer for the following periods: 1900–15, 1930–75, and 1990–95. The generally out-of-phase winter–summer relationship was also noted by Blasing, (1979, 1981), who applied a map-pattern correlation method to Northern Hemispheric SLP anomalies from 1900 to 1976 for the winter (December–February) and summer (July–August) seasons. Consistent with our Fig. 4, Blasing found the period from 1900 to 1920 to be characterized by a negative NAO pattern in summer and positive NAO pattern in winter. This seasonal pattern then flipped for the 1930–55 period, with summer (winter) having a positive (negative) NAO pattern. A primary conclusion from Fig. 4 is that low-frequency (decadal scales and longer) variations of the NAO have a strong seasonal dependence. All seasons can significantly amplify or diminish the low-frequency annual values of the NAOm.

d. Hemispheric teleconnections of NAO

As a final illustration of the seasonal dependence of the NAO, Fig. 5 shows spatial correlation fields between hemispheric SLP (reanalysis data) and two NAO indices: NAOm and W_H. The comparisons are shown for March, May, July, and November. These months particularly illustrate the seasonal differences in signal, strength, and associated teleconnections, and also highlight the large-scale spatial variability each index represents. The NAOm index, which has nodes mostly clustered in the central Atlantic basin (Fig. 1), generally has correlation maxima in the central North Atlantic, while W_H generally has correlation maxima along the eastern Atlantic basin rim. This difference is particularly evident for March (Figs. 5a,b). The subpolar correlation node of the March NAOm dipole is the strongest and broadest of all calendar months and extends deep into the Arctic. NAOm's basinwide subtropical correlation feature for March has teleconnections over the Balkans, northwestern Pacific basin, and western Mexico. The W_H is more strongly teleconnected to northern Africa and the Mediterranean region during March.

In May, the NAOm index again shows a much stronger and broader internode correlation signal in the central North Atlantic (Fig. 5c), while W_H index has a more diffuse pattern over the eastern Atlantic (Fig. 5d). The NAOm also exhibits a strong teleconnection to Mexico and the southwestern United States in May, when a positive correlation feature is located over the area of the July–August ''monsoon'' of the U.S. Southwest. The interannual variability of this monsoon has been linked to the late spring circulation over Mexico and the southwestern United States (Higgins et al. 1998). In July, the NAOm (Fig. 5e) has its westernmost nodal locations for the year, with a strong correlation dipole over the western Atlantic, but of smaller spatial scale than in late winter (Fig. 5a). In contrast, the July correlation field based on W_H contains no coherent dipole (Fig. 5f) since the subtropical high region lacks an organized area of significant correlation. The July and May teleconnection fields for NAOm also have a greater penetration into the Arctic than W_H. The same is true for the NAOm teleconnection fields in April and June (not shown).

Finally, the November teleconnection field for NAOm (Fig. 5g) has the same general character as in May, except that the November pattern does not include teleconnections with Mexico and the southwestern United States. In both May and November, the subtropical NAOm node is centered 2000–3000 km west of Lisbon. Figures 5g,h indicate that W_H better captures the strength of zonal flow into Europe for November, while NAOm more clearly reveals the zonality of flow over eastern North America and the western North Atlantic, including the Gulf Stream.

4. Conclusions

The results presented here show that the NAO is a year-round mode of atmospheric variability, with its nodes migrating seasonally in a manner that limits the utility of traditional station-based NAO indices in nonwinter months. The subtropical NAO node, in particular, shifts westward and northward into the central North Atlantic as the annual cycle passes through spring and summer. When tracked by our new ''mobile index'' (NAOm) that varies by calendar month, the NAO nodal correlations are only slightly weaker in the summer half year than in the winter half year. The mobile index further indicates that the NAO is strongest in March, followed by February and January.

The 1873–1997 time series of annual values of the NAOm shows that the most negative phase of the NAO occurred during the 1870s and 1880s. Because this phase resulted primarily from negative values during summer and autumn, it has not been apparent in previously published time series of the NAO (Hurrell 1995). Possible associations with ocean and sea ice conditions during this period have yet to be explored. During the first two decades of the twentieth century, spring and summer separately reinforced positive wintertime peaks of the NAO, so the time series of year-round and winter-only indices are similar during this period. The

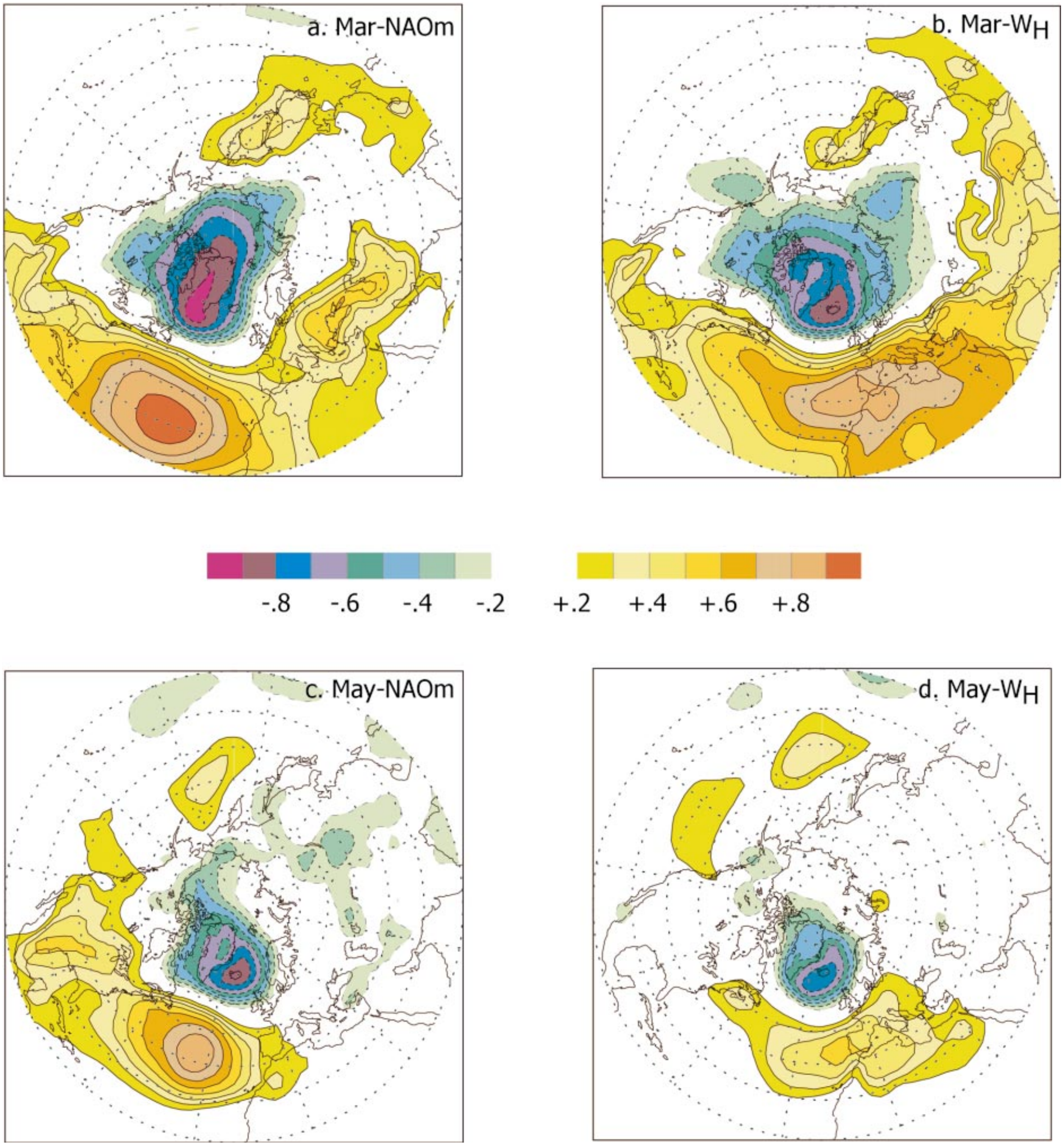


FIG. 5. Spatial correlation (1948–97) of reanalysis SLP with NAO indices (NAOm and W_H) for selected months: (a) Mar SLP and NAOm; (b) Mar SLP and W_H; (c) May SLP and NAOm; (d) May SLP and W_H; (e) Jul SLP and NAOm; (f) Jul SLP and W_H; (g) Nov SLP and NAOm; (h) Nov SLP and W_H. The 99% (95%) significance level is 0.36 (0.28).

generally negative NAO of the 1960s, and the increase to positive values in the 1980s and 1990s, are primarily due to winter patterns. Because these winter NAO characteristics have been partially offset by variations of the opposite sign during summer, the recent NAO increase is not as strong in the annual time series as in the winter-only time series. A potential caveat concerning the temporal NAO variation is that its character may change from one multidecadal period to another, thereby confounding representations of the NAO by seasonally varying nodes as well as by fixed-station indices and EOFs.

Monthly hemispheric teleconnection maps show that the broader associations of the NAO, particularly over

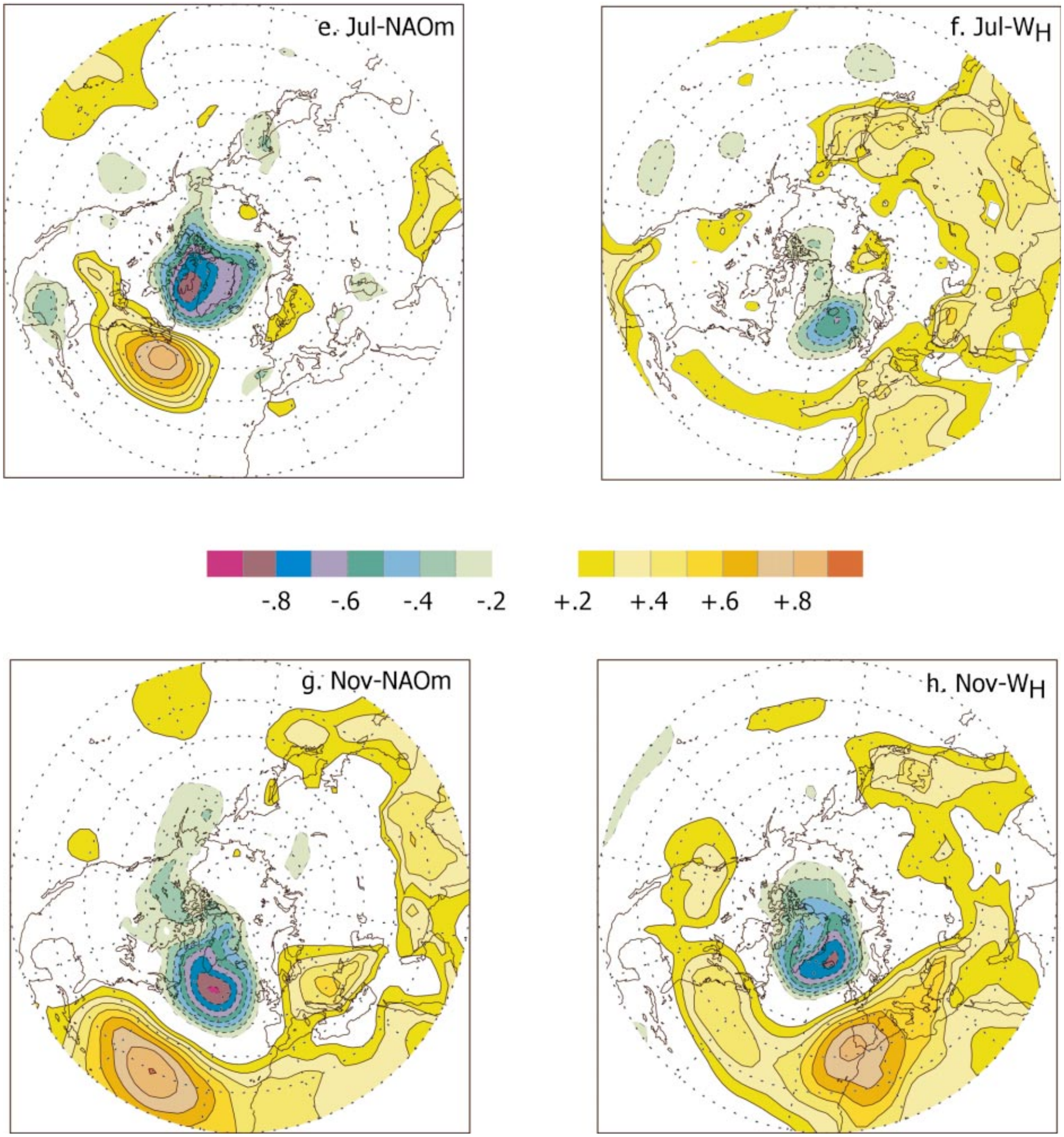


Fig. 5. (*Continued*)

the western North Atlantic and eastern North America, are more apparent during spring–summer–autumn when the new NAOm index is used. The penetration of NAO teleconnection fields into the Arctic during April–July was also greater when the mobile index was used in place of the geographically fixed Iceland–Lisbon index. This new NAOm index may therefore be useful in placing the NAO into the broader context of hemispheric-scale variability.

The seasonal inclusivity of the NAOm index may also permit more comprehensive assessments of climate model simulations of the NAO. To date, such assessments have not generally included seasonality (e.g., Mehta et al. 2000). Finally, to the extent that the character of surface boundary forcing varies seasonally, the NAOm may be a useful tool in diagnosing the seasonality of ocean–atmosphere coupling in the North Atlantic.

Acknowledgments. This work was supported by the Climate Dynamics Program of the National Science Foundation (Grant ATM-9905924), the NOAA Office of Global Programs (Grant NA67RJ0150), and the U.S. Agency for International Development (Grant 608-G-00-94-00007). The reanalysis data were obtained from the NOAA–CIRES Climate Diagnostics Center. We thank Jeff Rogers for providing helpful comments on the original manuscript (as a formal peer reviewer) and for the use of his NAO index. The comments of two other reviewers also helped improve the manuscript.

REFERENCES

Barnston, A. G., and R. E. Livezey, 1987: Classification, seasonality, and persistence of low-frequency atmospheric circulation patterns. *Mon. Wea. Rev.,* **115,** 1083–1126.

Barry, R. G., and A. H. Perry, 1973: *Synoptic Climatology: Methods and Applications.* Methuen and Co. Ltd., 555 pp.

Blasing, T. J., 1979: Map pattern classification at a prescribed level of generality. Preprints, *Sixth Conf. on Probability and Statistics in the Atmospheric Sciences,* Banff, Alberta, Canada, Amer. Meteor. Soc., 118–125.

——, 1981: Characteristic anomaly patterns of summer sea-level pressure for the Northern Hemisphere. *Tellus,* **33,** 428–437.

Cheng, X., G. Nitsche, and J. M. Wallace, 1995: Robustness of low-frequency circulation patterns derived from EOF and rotated EOF analyses. *J. Climate,* **8,** 1709–1713.

Corti, S., F. Molteni, and T. N. Palmer, 1999: Signature of recent climate change in the frequencies of natural atmospheric circulation regimes. *Nature,* **398,** 799–802.

Davis, R. E., B. P. Hayden, D. A. Gay, W. L. Phillips, and G. V. Jones, 1997: The North Atlantic subtropical anticyclone. *J. Climate,* **10,** 728–744.

Deser, C., and M. L. Blackmon, 1993: Surface climate variations over the North Atlantic Ocean during winter: 1900–1989. *J. Climate,* **6,** 1743–1753.

Dickson, R. R., 1997: From the Labrador Sea to global change. *Nature,* **386,** 649–650.

——, and J. Namias, 1976: North American influences on the circulation and climate of the North Atlantic sector. *Mon. Wea. Rev.,* **104,** 1256–1265.

Duchon, C. E., 1979: Lanczos filtering in one and two dimensions. *J. Appl. Meteor.,* **18,** 1016–1022.

Glowienka-Hense, R., 1990: The North Atlantic Oscillation in the Atlantic-European SLP. *Tellus,* **42A,** 497–507.

Hastenrath, S., 1991: *Climate Dynamics of the Tropics.* Kluwer Academic Publishers, 488 pp.

Higgins, R. W., K. C. Mo, and Y. Yao, 1998: Interannual variability of the U.S. summer precipitation regime with emphasis on the southwestern monsoon. *J. Climate,* **11,** 2582–2606.

Hilmer, M., and T. Jung, 2000: Evidence for a recent change in the link between the North Atlantic Oscillation and Arctic sea ice export. *Geophys. Res. Lett.,* **27,** 989–992.

Hurrell, J. W., 1995: Decadal trends in the North Atlantic Oscillation: regional temperatures and precipitation. *Science,* **269,** 676–679.

——, and H. van Loon, 1997: Decadal variations in climate associated with the North Atlantic Oscillation. *Climatic Change,* **36,** 301–326.

Jones, P. D., 1987: The early twentieth century Arctic High—Fact or fiction? *Climate Dyn.,* **1,** 63–75.

Kalnay, E., and Coauthors, 1996: The NCEP/NCAR 40-Year Reanalysis Project. *Bull. Amer. Meteor. Soc.,* **77,** 437–471.

Kushnir, Y., 1994: Interdecadal variations in North Atlantic sea surface temperature and associated atmospheric conditions. *J. Climate,* **7,** 141–157.

Lamb, H. H., 1972: *Climate: Past, Present and Future.* Vol. 1. Methuen and Co., Ltd., 613 pp.

Lamb, P. J., and R. A. Peppler, 1987: North Atlantic Oscillation: Concept and application. *Bull. Amer. Meteor. Soc.,* **68,** 1218–1225.

Mächel, H., A. K. Kapala, and H. Flohn, 1998: Behaviour of the centres of action above the Atlantic since 1881. Part I: Characteristics of seasonal and interannual variability. *Int. J. Climatol.,* **18,** 1–22.

Mehta, V. M., M. J. Suarez, J. V. Manganello, and T. L. Delworth, 2000: Oceanic influence on the North Atlantic Oscillation and associated Northern Hemispheric climate variations: 1959–1993. *Geophys. Res. Lett.,* **27,** 121–124.

Peixoto, J. P., and A. H. Oort, 1992: *Physics of Climate.* American Institute of Physics, 520 pp.

Rodwell, M. J., D. P. Rowell, and C. K. Folland, 1999: Oceanic forcing of the wintertime North Atlantic Oscillation and European climate. *Nature,* **398,** 320–323.

Rogers, J. C., 1984: The association between the North Atlantic Oscillation and the Southern Oscillation in the Northern Hemisphere. *Mon. Wea. Rev.,* **112,** 1999–2015.

Sahsamanoglou, H. S., 1990: A contribution to the study of action centres in the North Atlantic. *Int. J. Climatol.,* **10,** 247–261.

Thompson, D. W. J., and J. M. Wallace, 1998: The Arctic Oscillation signature in the wintertime geopotential height and temperature fields. *Geophys. Res. Lett.,* **25,** 1297–1300.

——, ——, and G. C. Hegerl, 2000: Annual modes in the extratropical circulation. Part II: Trends. *J. Climate,* **13,** 1018–1036.

Trenberth, K. E., and D. A. Paolino Jr., 1980: The Northern Hemisphere sea-level pressure data set: Trends, errors and discontinuities. *Mon. Wea. Rev.,* **108,** 855–872.

van Loon, H., and J. C. Rogers, 1978: The seesaw in winter temperatures between Greenland and northern Europe. Part I: General description. *Mon. Wea. Rev.,* **106,** 296–310.

Walker, G. T., and E. W. Bliss, 1932: World Weather V. *Memoirs Roy. Meteor. Soc.,* **44,** 53–84.

Wallace, J. M., 2000: North Atlantic Oscillation/annular mode: Two paradigms–one phenomenon. *Quart. J. Roy. Meteor. Soc.,* **126,** 791–805.

——, and D. S. Gutzler, 1981: Teleconnections in the geopotential height field during the Northern Hemisphere winter. *Mon. Wea. Rev.,* **109,** 784–812.

Ward, M. N., P. J. Lamb, M. El Hamly, R. Sebbari, and D. H. Portis, 1999: Climate variability in Northern Africa: Understanding droughts in the Sahel and the Maghreb. *Beyond El Niño: Decadal Variability in the Climate System,* A. Navarra, Ed., Springer Verlag, 119–140.