Temporal Variability in the Expression of the Arctic Oscillation in the North Pacific

HONGXU ZHAO* AND G. W. K. MOORE

Department of Physics, University of Toronto, Toronto, Ontario, Canada

(Manuscript received 30 April 2008, in final form 11 November 2008)

ABSTRACT

Although the Arctic Oscillation (AO) and North Atlantic Oscillation (NAO) have been identified as important modes of climate variability during the Northern Hemisphere (NH) winter, whether the AO or the NAO is more fundamental to the description of this variability, especially in the North Pacific, is still an open question. An important contributor to this uncertainty is the lack of knowledge of the low-frequency linkages between the North Atlantic and North Pacific Oceans. This paper explores the linkage between the two oceanic basins on interdecadal time scales using the sea level pressure (SLP) field during the twentieth century. In particular, it is shown that the winter mean SLP in the North Pacific was positively correlated with the sign of the NAO during the periods of 1925–50 and 1980–98, which resulted in the classical AO pattern being the dominant mode in the NH. In contrast, during the period of 1951–79, the winter mean SLP in the two basins was decoupled, resulting in a dominant mode that more closely resembled the NAO. Using paleoclimate reconstructions, it is also shown that this interdecadal variability in the North Pacific climate began around 1850, which is nominally considered to be the end of the Little Ice Age.

1. Introduction

The North Atlantic Oscillation (NAO) and the Arctic Oscillation (AO) are two of the most important modes characterizing climate variability especially during the boreal winter (Hurrell 1995; Thompson and Wallace 1998, hereafter TW98; Thompson and Wallace 2000; Wanner et al. 2001; Rigor et al. 2002). Both represent modes of variability associated with a meridional displacement of atmospheric pressure in the Northern Hemisphere (NH), with the NAO being more regionally focused in the North Atlantic region, while the AO has a more zonally symmetric “annular” structure. The NAO was first identified in the pioneering work of Walker as part of his investigation into predictability of the Indian monsoon (Walker and Bliss 1932). It is now conventional to define it in terms of the difference in sea level pressure (SLP) between its two centers of action, one near Iceland and the other near the Azores.

The AO is a more recent discovery identified as the leading mode in an empirical orthogonal function (EOF) decomposition of the NH winter (November–April) monthly mean SLP anomaly field (TW98; Thompson and Wallace 2001; Wallace and Thompson 2002, hereafter WT02). It is characterized by a center of action over the Arctic Ocean with centers of opposite sign over the North Atlantic and North Pacific Oceans. It has furthermore been suggested that the NAO is simply the regional manifestation of the AO in the North Atlantic (Wallace 2000; WT02). Along these lines, TW98 have argued that the AO is a more fundamental structure because it accounts for a substantially larger fraction of the variance of the NH surface air temperature than does the NAO.

In response to Thompson and Wallace’s assertion, a vigorous debate has taken place surrounding the issue of whether the AO or the NAO represents the main mode in NH wintertime tropospheric climate variability. For example, Ambaum et al. (2001) used the NH winter mean SLP field from the 15-yr European Centre for Medium-Range Weather Forecasts (ECMWF) Re-Analysis (ERA-15) over the period 1979–97 to examine the differences between the AO and NAO and suggested that the NAO and Pacific North–America (PNA) pattern (Barnston and Livezey 1987) rather the AO represent the more physically relevant and robust modes for describing NH tropospheric climate variability. Deser

* Current affiliation: Earth Sciences Sector, Natural Resources Canada, Ottawa, Ontario, Canada.

Corresponding author address: G. W. K. Moore, Department of Physics, University of Toronto, Toronto, ON M5S 1A7, Canada.
E-mail: gwk.moore@utoronto.ca

DOI: 10.1175/2008JCLI2611.1

© 2009 American Meteorological Society
used identical datasets and the same analysis period as those used by TW98 to investigate the correlations among the three centers of action of the AO. She argued that the AO is characterized predominantly by its Arctic center of action and that there exists only a weak positive correlation between the two centers of action over the North Atlantic and North Pacific. Dommengen and Latif (2002) provided an idealized example of how the superposition of geographically separated uncorrelated structures could result in a single mode through the application of an EOF decomposition.

WT02 used the monthly National Centers for Environmental Prediction (NCEP) reanalysis dataset over the period 1958–99 to rebut these arguments. In particular, they argued that the lack of correlation in the SLP field between the North Atlantic and North Pacific did not in and of itself preclude the applicability of the paradigm as to the predominance of the AO. They suggested that the lack of correlation between the Atlantic and Pacific centers of the AO was due to the presence of the PNA, which they defined as the second mode of the EOF decomposition of the NH winter SLP field. To support this contention, they compared one-point correlation maps based on SLP field with partial (residual) correlation maps created by removing the second leading EOF mode from the original SLP field. Through this technique, they showed that there was a positive correlation between the SLP fields within the North Atlantic and North Pacific centers of the AO.

Implicit in all of these arguments is the assumption as to the stationarity of the spatial patterns, be it the AO, the NAO, or the PNA, that describes climate variability in the NH. As a result, many of the studies quoted above have used differing time periods, often dictated by availability of certain datasets, in their analyses. For example, TW98 and Deser (2000) used the time period 1947–97 in their analysis, while WT02 used the period 1958–99. In contrast, Ambaum et al. (2001) used the considerably shorter period of 1979–97 in their analysis. In addition, there has been an inconsistency in the definition of the base fields upon which the various EOF decompositions have been applied. For example, TW98, Deser (2000), and WT02 used monthly mean fields during the winter, defined as the period from November through March. In contrast, Ambaum et al. (2001) used winter mean fields, defined as the mean of the months of December–February, in their analysis. The use of monthly mean fields clearly includes information on both intra-annual and interannual climate variability, while the use of winter mean fields filters out the former and as a result only includes information on the latter (Deser 2000).

We show in Fig. 1 the impact that differing time periods and differing definitions of the base fields have on the EOF decomposition of the NH winter SLP field. Please refer to section 2 for information on the techniques used to generate this figure. In the case of winter mean, over the periods 1947–97 and 1958–97 (Figs. 1a,b), the leading EOF does not show a Pacific center of action, while that over the period 1979–97 (Fig. 1c) does show such a center. If one uses monthly mean fields, the Pacific center of action appears for all the three periods but with increasing areal coverage (Figs. 1d–f). This simple example suggests that the spatial structures that describe variability in the wintertime SLP field may vary on interdecadal time scales in a fashion that depends on whether winter mean or monthly mean fields during the winter are used. This temporal and spatial variability may partially explain some of the contradictory results that have fuelled the debate over the relative importance of the AO and the NAO in describing boreal winter climate variability in the North Pacific.

To provide further evidence of the intra-annual variability in the sea level pressure field, we show in Fig. 2 the climatological monthly mean SLP from November to March as well as the climatological winter mean SLP over the same months. It can be seen that three main centers of action (Aleutian low, Icelandic low, and Siberian high) have seasonal variability. For example, the Aleutian low is weaker in November and March but stronger in January and February. As a result, the Aleutian low deepens and extends southwards from November to January, and it weakens and shrinks from January to March. Because of this seasonal variability, the AO-like EOF pattern in Figs. 1d–f includes both intra-annual and interannual variability, and it is our assertion that the EOF1 of the monthly mean SLP fields cannot represent the winter mean variability in the SLP field.

The North Pacific region has been identified as an area that experiences variability on interdecadal time scales. Minobe (1997), using a reconstructed air temperature time series based on North American tree-ring data, identified variability with a periodicity of 50–70 yr in the region over the past few centuries. The leading EOF of the North Pacific sea surface temperature (SST) field, referred to as the Pacific (inter) decadal oscillation (PDO), has been shown to exhibit variability on these time scales as well (Mantua et al. 1997; Minobe 1997). The spatial pattern of the PDO is similar to that of the El Niño–Southern Oscillation (ENSO) but with a more pronounced center of action in the central North Pacific near 40°N and 170°W (Zhang et al. 1997; Overland et al. 2000). The existence of the PDO results in fundamental changes in the mean state of the regional climate,
FIG. 1. The leading EOF (EOF1) of the winter (JFM) mean SLP northward of 20°N over the periods (a) 1947–98, (b) 1958–98, and (c) 1979–98. (d)–(f) The same as (a)–(c) except for monthly mean (JFM) SLP. Note that the EOFs have been multiplied by 100.
Fig. 2. The climatological monthly mean SLP (hPa) northward of 20°N over the period 1900–98 for (a) November, (b) December, (c) January, (d) February, and (e) March; (f) the winter mean SLP (hPa) averaged from November to March.
referred to as the regime shifts (Minobe 1997). During the twentieth century, shifts associated with the PDO have been identified to occur around 1925, 1948, and 1977 (Mantua et al. 1997; Minobe 1997; Minobe and Mantua 1999; Deser et al. 2004). Overland et al. (2000) suggested that, in the North Pacific, the coupled ocean–atmosphere system is less sensitive to external forcing when it is within one of the stable states of the PDO. As a result, the climate system in the North Pacific may involve nonlinear coupling with other regions, such as the North Atlantic, when undergoing regime shifts. Zhao and Moore (2004) showed that there does exist a nonlinear expression of the NAO in the North Pacific that is amplitude but not sign dependent over the period of 1948–2000. Within this period, the North Pacific climate experienced a regime transition around the late 1970s. Therefore, their results may partially be a response to the regime shift in the North Pacific. Schwing et al. (2003) also identified a nonlinear phenomenon on decadal time scales between the NAO and PDO with negative correlation prior to the late 1950s but positive correlation after the transition period.

In this paper, we will show that there exists variability on interdecadal time scales in the Pacific center of action in the AO mode, which is associated with regime shifts of the PDO. As we shall show, this nonstationarity can result in a finding as to the dominance of either the AO or the NAO as a paradigm for understanding NH tropospheric climate variability.

2. Data and methods

We will make use of the First Hadley Centre Sea Level Pressure dataset (HadSLP1), produced by the Met Office (Basnett and Parker 1997). This dataset makes use of all available SLP data over both land and ocean to produce an internally consistent time series of the global monthly mean SLP field over the period 1871–1998 on a 5° latitude–longitude grid. There is concern as to the lack of observations over the North Pacific Ocean prior to the early part of the twentieth century (Trenberth and Hurrell 1994). As a result, we restrict our attention to the period after 1900.

We will use the Hurrell (1995) NAO index (NAOI), which uses the difference of the normalized Lisbon, Portugal, and the Stykkisholmur, Iceland, SLP data. This particular index is widely used in studies of climate variability associated with the NAO (Wanner et al. 2001). This NAOI is available on a monthly mean basis starting in 1821. To describe climate variability in the North Pacific, we make use of the Aleutian low index (ALI). It is an area-weighted average of the monthly SLP over the region 30°–65°N, 160°E–140°W. More information on this index, originally referred to as the North Pacific index, and its importance to North Pacific and North American climate is detailed in Trenberth and Hurrell (1994). The AO index (AOI) is the principal component of leading EOF of the monthly mean winter SLP north of 20°N. It is available from 1899 onward. To address the concern that the AOI obtained from EOF analysis may constrain the results, another index of the AO (AOI2), defined as the normalized difference in zonal-averaged SLP anomalies between 35° and 65°N, will also be used (Li and Wang 2003). With regard to an index of the PDO (PDOI), we used the principal component of the leading EOF of the monthly mean SST field poleward of 20°N for the period 1900–2004 (Mantua et al. 1997). For a longer-term perspective, we also use the reconstructed indexes of the PDO (D’Arrigo et al. 2001) and the NAO (Luterbacher et al. 2002) over the period 1700–1997.

EOF decomposition is a commonly employed data compression technique used to extract modes or spatial patterns that maximize the amount of variance explained in a given dataset (Richman 1986). The EOFs are usually defined as the eigenvectors of the covariance matrix of a given dataset (Preisendorfer et al. 1988). The EOFs provide for both spatial patterns and associated time series, which is referred to as expansion coefficients or principal components (PCs). Because the EOFs do not necessarily represent physical modes, care must be used in interpretation of such analyses (Richman 1986; Ambaum et al. 2001). Dommenger and Latif (2002) argued that EOF analyses can have problems in identifying the dominant centers of action and recommended care in interpreting the results of these analyses. They suggested a strategy for identifying physical modes of climate variability through the use of multiple techniques, including EOF decompositions as well as regression analysis. In this study, following their recommendations (Dommenger and Latif 2002), both EOF decomposition and regression analysis are applied. We used the Student’s t test to assess the statistical significance of the regressions against meteorological fields (von Storch and Zwiers 1999).

The correlation coefficient (CC) and moving window correlation (MWC) are simple ways to test the existence of a linear relationship between two time series, and the latter can identify the temporal evolution of the relationship, which may be associated with regime transitions. The statistical significance of the MWC was assessed using a resampling technique that uses red noise synthetic time series with the same statistical properties as the underlying time series (Allen and Smith 1996; Biondi et al. 2001). Concerns have been recently raised (Gershunov et al. 2001) regarding the degree to which interdecadal variability in the relationship...
between climate indexes identified using MWC necessarily has a physical basis or may just be the result of stochastic processes. The significance test, which is based on such a resampling technique, provides additional support for a physical basis for the variability that we identify.

Please note that we will use the winter mean fields/indexes defined as the mean of three months: January, February, and March (JFM) throughout the study. As discussed above, this choice clearly focuses attention on interannual climate variability. Other authors have used monthly mean fields for the winter months in their analyses and as a result have included both intra-annual and interannual variability. As shown in Fig. 1, this choice has a significant impact on the EOF decomposition. For the reasons outlined above, we feel that our approach is the most relevant for the study of low-frequency climate variability. We also tested the robustness of our results with respect to the choice of months to include in the winter mean. We found that including any continuous subset of months from November through March did not change our results in any significant way.

3. Results

a. Temporal evolution of the correlation between the NAOI and ALI

Figure 3 shows the MWCs between the winter mean NAOI and ALI as well as the NAOI and PDOI over the period of 1900–2000 with a 21-yr moving window. Other interdecadal window lengths produced similar results. The statistical significance of the MWCs was assessed through the generation of 1000 synthetic indexes for each of the NAOI, ALI, and PDOI that had the same red noise characteristics as the underlying time series (Rudnick and Davis 2003). These synthetic indexes were then used to estimate the probability distribution function for the various MWCs, and from these functions their statistical significance was assessed. As discussed above, we believe that this approach addresses concerns raised by Gershunov et al. (2001).

It can be seen that the correlation between the NAOI with the ALI undergoes changes in both magnitude and sign throughout the period. In particular, the intervals between 1925 and 1950 and 1980 and 1990 are ones where these MWCs are both positive and for the most part statistically significant at close to the 90% level. In contrast, the interval between 1950 and 1979 is one where the MWC between the ALI and NAOI is negative, indicating a completely opposite coupling between the two indexes over this period. The MWC between the NAOI and the inverted PDOI is also shown in this figure and it displays similar behavior to that between the NAOI and ALI. It is noted that there is some variability in the region of maximum correlation in the North Pacific SLP coupling with the NAOI over these subperiods. Hence, it is possible to increase the magnitude of the correlation by defining new subindexes that reflect this variability.

As a further test as to the robustness of the three subperiods identified in Fig. 3 (double-headed arrows), we show in Table 1 the CCs between the four instrumental indexes, AOI, ALI, NAOI, and inverted PDOI, for these three subperiods. We see that, for the subperiods 1925–50 and 1980–2000, the AOI is positively correlated at a confidence level of 90% with the ALI. In contrast, for the subperiod 1950–79, the correlation is not robust between the 2 indexes. A similar pattern arises with respect to the correlation between the AOI and the inverted PDOI. The correlation between the NAOI and the ALI is not statistically significant for any of these subperiods. There is however a change in the sign of the correlation between the first and second and the second and third subperiods consistent with the results from Fig. 3. To assess the significance of this behavior, we used our synthetic time series to test the probability of changes in the sign and magnitude of the correlation between the ALI and NAOI as identified in Table 1. The test shows that the probability of observing the two changes in the magnitude and sign of the correlation between the ALI and NAOI identified in Table 1 was less than 5%, indicating statistical significance at the 95% level. In contrast, there is a statistically significant correlation between the NAOI and the inverted PDOI during the first and third subperiods. The significance test on the change in sign of the correlations between the PDOI and NAOI also shows 95% significance level for both transitions. With respect to the correlation between the AOI and the NAOI, it is positive and statistically significant over all three subperiods. Similar results were obtained with the AOI2 index of the Arctic Oscillation.

b. Temporal variability in the spatial patterns of the leading EOFs of the winter SLP in the NH during the twentieth century

Given the results presented in Fig. 3 and Table 1, it is evident that there is temporal variability in the nature of the correlations between major NH climate indexes during the twentieth century with changes in sign occurring around 1950 and 1980. We seek to identify the spatial imprint of this variability by presenting the two leading EOFs of the SLP field for these different periods of time.
The first two EOFs (EOF1 and EOF2) of the winter mean SLP field are shown in Figs. 4 and 5, respectively, over the full period of 1900–98 as well as over the three subperiods of 1925–50, 1951–79, and 1980–98. For the full period, 1900–98, EOF1 contained a small center of action in the North Pacific (Fig. 4a). Looking at the three subperiods, we see that there exists apparent temporal variability in the North Pacific region with a positive center of action that leads to an AO-like pattern as the expression of EOF1 during the subperiods 1925–50 (Fig. 4b) and 1980–98 (Fig. 4d), while the lack of such center of action in the North Pacific leads to a NAO-like pattern during 1951–79 (Fig. 4c). It is also noted that there is some variability in the locations of the various centers of action during the various time periods. Figure 5 shows the spatial structure of EOF2 over the same periods as in Fig. 4. The EOF2 is restricted to the North Pacific and resembles the Aleutian low, except over the subperiod 1950–79 when there is a larger center of action in northern Europe than in the North Pacific. Note that EOF2 is only independent of EOF1 over the same subperiod. In summary, over the subperiods 1925–50 and 1980–98, the leading EOF closely resembles the conventional AO pattern. In contrast, the lack of a North Pacific center of action during the subperiod 1951–79 results in a leading EOF pattern that more closely resembles the NAO rather than the AO. Over the full period of 1900–98, the EOF1 only presents a small North Pacific center.

To confirm the extent to which the leading EOFs convey information on temporal variability associated with the AO, the NAO, and the Aleutian low, the CCs between the principal components of the leading two EOFs over the period 1900–98 with the AOI, NAOI, and ALI are shown in Table 2. Results are shown for the full period 1900–98 as well as the three subperiods identified above. Over the entire period, it can be seen

<table>
<thead>
<tr>
<th></th>
<th></th>
<th></th>
<th></th>
</tr>
</thead>
<tbody>
<tr>
<td>AOI vs ALI</td>
<td>0.40</td>
<td>0.19</td>
<td>0.36</td>
</tr>
<tr>
<td>AOI vs PDOI</td>
<td>0.51</td>
<td>0.21</td>
<td>0.59</td>
</tr>
<tr>
<td>NAOI vs ALI</td>
<td>0.25</td>
<td>-0.21</td>
<td>0.19</td>
</tr>
<tr>
<td>NAOI vs PDOI</td>
<td>0.51</td>
<td>-0.16</td>
<td>0.45</td>
</tr>
<tr>
<td>NAOI vs AOI</td>
<td>0.85</td>
<td>0.74</td>
<td>0.86</td>
</tr>
<tr>
<td>PDOI vs ALI</td>
<td>-0.56</td>
<td>-0.79</td>
<td>-0.57</td>
</tr>
</tbody>
</table>
that there is a statistically significant correlation between the EOF1 and the AOI, NAOI, and ALI. The magnitude of the correlation decreases from close to 1 for the AOI to less than 0.3 for the ALI. In contrast, over the full period there is a statistically significant correlation between the EOF2 and the ALI but not with either the AOI or NAOI. Similar results are obtained with respect to the first subperiod, 1925–50. With respect to the second subperiod, 1951–79, there exist correlations of a similar magnitude between EOF1 and both the AOI and NAOI as occurred for the entire period and the first subperiod. However, the correlation between the EOF1 and the ALI is smaller in magnitude and no longer statistically significant. With regards to the EOF2, we note that a statistically significant correlation exits between it and both the ALI and the NAOI for this time period. The correlation with the NAOI is in marked difference to what occurred for the entire period and the first subperiod. Correlations for the third subperiod, 1980–98, are similar to what occurred for the entire period as well as the first subperiod with the exception that the magnitude of the correlation between the EOF1 and the ALI is just under the threshold for statistical significance at the 90% confidence level. It

FIG. 4. EOF1 of the winter (JFM) mean SLP northward of 20°N over (a) 1900–98, (b) 1925–50, (c) 1951–79, and (d) 1980–98. Note that the EOFs have been multiplied by 100.
should be noted that we also performed the above correlations but using the principal components of the EOFs that are calculated for the respective subperiods. The results obtained are similar to what was presented in Table 2.

c. Temporal variability in regression patterns of the winter SLP in the NH during the twentieth century

In sections 3a and 3b, we have identified interdecadal temporal variability in the leading patterns of NH climate variability through the use of moving window

![Fig. 5. Same as in Fig. 4, but for EOF2.](image)

| Table 2. Correlation coefficients of instrumental indexes with EOF1 and EOF2 for different time periods. The bold values are significant at 90% confidence using the Student’s t test. |
|-----------------|-------|-----|-----|-----|-----|
| AOI vs EOF1     | 0.97  | 0.96 | 0.98 | 0.99 |
| NAOI vs EOF1    | 0.84  | 0.85 | 0.75 | 0.87 |
| ALI vs EOF1     | 0.26  | 0.37 | 0.14 | 0.33 |
| AOI vs EOF2     | 0.02  | 0.14 | 0.01 | 0.09 |
| NAOI vs EOF2    | −0.15 | 0.06 | −0.38 | −0.15 |
| ALI vs EOF2     | 0.78  | 0.86 | 0.90 | 0.90 |
correlations and EOF analysis. Following Dommenget and Latif (2002), we will confirm these results with regressions of the winter mean SLP field against the winter mean NAOI (Fig. 6), AOI (Fig. 7), AOI2 (not shown), ALI (Fig. 8), and PDOI (Fig. 9). In each figure, results are shown for the full period of 1900–98 as well as over the three subperiods of 1925–50, 1951–79, and 1980–98. The statistical significance of these patterns was assessed using the Student’s t test (von Storch and Zwiers 1999).

From Fig. 6, it can be seen that the dipolar structure of the regression pattern is similar to that usually associated with the NAO, especially over the full period (Fig. 6a). However, in the North Pacific there are positive correlations over 1925–50 (Fig. 6b) and 1980–98 (Fig. 6d) that result in a more annular-like structure. Over the subperiod 1951–79 (Fig. 6c), we see that the signal in the Pacific is weak and in phase with that of Arctic center but out of phase with that of the Atlantic center. Furthermore, the centers of action in the two oceanic basins are displaced toward the American continent during 1925–50 (Fig. 6b) but toward the Eurasian continent during the most recent subperiod (Fig. 6d), which is also reflected in the magnitude of action centers in the leading EOFs (Fig. 4), verifying the existence of some variability in the region of maximum correlation in the North Pacific SLP with NAOI.

Figure 7 shows the regression of the winter mean SLP field against the winter mean AOI over the four time periods under investigation. The results are consistent with the results of the EOF analysis (Fig. 4) in that a well-defined AO-like pattern with three centers of action is present during the full period as well as the first (1925–50) and the third (1980–98) subperiods. In contrast, during the second subperiod (1950–79), the regression has a more NAO-like structure. For completeness, the regressions of the winter mean SLP against the winter mean AOI2 was calculated (not shown) and a comparison with Fig. 7 shows the same general patterns over the various time periods. It is worth mentioning that the AO-like pattern over the full period should result from the contributions of those over the subperiods of 1925–50 and 1980–98.

Figure 8 shows the regression of the winter mean SLP field against the winter mean ALI. A statistically significant center of action exists in the North Pacific during all the time periods, indicating that, as expected, the ALI is highly correlated with the Aleutian low. These regression patterns are to some extent similar to the second EOF in Fig. 5, except during 1951–79. With respect to the North Atlantic, we see that there exists a statistically significant signal of the ALI in the region, except during the second subperiod, 1951–79. The smaller statistically significant area in the North Atlantic for the first subperiod results from variability in the region of maximum correlation in the North Pacific SLP with the NAOI. These results are again consistent with the regression patterns in Figs. 6, 7 as well as EOF analyses in Figs. 4, 5.

We conclude this section with Fig. 9, which shows the regression of the winter mean SLP field against the winter mean PDOI. In agreement with previous results, the PDOI regression patterns over 1925–50 and 1980–98 have an AO-like annular structure both in correlation coefficients of a similar sign and in statistically significant area in the North Pacific and North Atlantic Oceans. In contrast, there is a weakened annular structure over the full period as well as the second subperiod of 1951–79.

d. Temporal variability in the reconstructed NAOI and PDOI time series

The lack of SLP observations over the North Pacific prior to the early years of the twentieth century makes it difficult to place the results obtained above in a long-term context (Trenberth and Hurrell 1994). The transitions that we have identified are approximately coincident with regime shifts of the PDO since the 1920s (Minobe 1997). Earlier regime shifts of the PDO have been identified (Gedalof and Smith 2001) but unfortunately cannot be studied using the available global SLP datasets. Reconstructions of the PDO and the NAO are, however, available for at least the past 300 yr (D’Arrigo et al. 2001; Luterbacher et al. 2002), and we can use them to consider the behavior of the PDO in determining the spatial pattern of NH winter SLP variability in the preinstrumental period. Please note that the reconstructed PDOI is based exclusively on tree-ring data and primarily represents the spring and summer seasons (D’Arrigo et al. 2001). However, because of the persistence of SST anomalies, it is found that the reconstructed PDOI is also significantly correlated to the winter instrumental PDOI (D’Arrigo et al. 2001). It should be emphasized that other reconstructions of the PDO (Biondi et al. 2001; Gedalof and Smith 2001) that make use of different collections of tree rings do not agree with the PDOI of D’Arrigo et al. (2001) in all aspects and the reasons for this are not clear at this time.

Figure 10 presents the 21-yr MWC between the reconstructed winter mean PDOI and the winter mean NAOI for the period 1700–2000. In addition, estimates of the significance of the MWC as determined by the resampling technique are also provided. Superimposed on the long-term MWC curve is the corresponding curve generated with the instrumental indexes. To facilitate comparison with Fig. 3, we have inverted the
PDO indexes prior to the application of the MWC. As one might expect, both MWCs agree well during the instrumental time period. In addition, we see that, for windows centered around 1900, there exists a statistically significant correlation between the reconstructed PDOI and NAOI that is similar to that which existed between 1951 and 1979. Around 1850, there appears to be a fundamental change in the nature of the correlation between the two basins. Prior to this time, the correlation between the two indexes undergoes frequent reversals in the sign with only a small number of windows achieving the 90% significance level. In contrast, after 1850, there are less frequent reversals in the sign of the correlation with a large number of windows exceeding the 90% significance threshold indicating interdecadal variability.

4. Summary and discussion

In this study, we have identified temporal variability in the Pacific center of action in the AO mode that is related to the sign of the correlation between the NAOI and the ALI on a multidecadal time scale. Periods of time when this correlation was positive, such as 1925–50

**Fig. 6.** Regressions of the winter (JFM) mean SLP at every grid point against the winter (JFM) NAOI over the periods (a) 1900–98, (b) 1925–50, (c) 1951–79, and (d) 1980–98. The shaded regions are statistically significant at 90% level using the Student’s t test.
and 1980–98, are characterized by having a classical AO-like pattern as the leading EOF of the winter mean SLP field. In contrast, the period from 1951 to 1979, in which the correlation between the NAOI and the ALI was negative, has a NAO-like pattern as the leading EOF. The results of the EOF analysis were confirmed through the use of regressions of the winter mean SLP field against indexes for the Aleutian low, the NAO, and the AO. The timing of the reversals in the sign of the correlation between the NAOI and the ALI during the twentieth century is similar to those identified for regime shifts associated with the PDO, that is, 1925, 1947, and 1977. It is therefore natural to suggest that the variability we observe in the leading pattern of winter mean SLP field in the North Pacific is related to the low-frequency variability in the SST field in the region as expressed in the PDO. Indeed, we have replicated the multidecadal reversal in the correlation between the NAOI and the ALI through the use of the NAOI and the PDOI.

Our results suggest that a nonlinearity exists between the climate of North Pacific and the North Atlantic that is a function of the sign of the PDO. When the PDO is in its negative phase, the SLP field in the region is tightly coupled with the PDO resulting in little or no correlation with the North Atlantic resulting in a leading mode of hemispheric SLP variability that resembles the NAO. In contrast, when the PDO is in its positive phase, the coupling of the SLP field in the North Pacific is weaker, resulting in a leading mode of hemispheric SLP variability that resembles the AO with a tight coupling between the North Pacific and North Atlantic.
Such a time-dependent nonlinearity would partially explain the seemingly contradictory conclusions reached by a number of studies (TW98; Deser 2000; Ambaum et al. 2001; WT02) as to whether the AO or the NAO represents the main mode in NH wintertime tropospheric climate variability. Although TW98 and Deser (2000) used the identical dataset, it includes a regime shift in the late 1970s. The selection of a time interval for such a determination that includes a regime shift of the PDO would perhaps result in the mixing of both the AO-like and NAO-like patterns with the conclusion reached dependent on details of the analysis method. This indeterminacy is in contrast to the clear identification of either the NAO or the AO as the leading mode of SLP variability once time periods are selected that do not involve a regime shift of the PDO.

Our results are consistent with those of Minobe and Mantua (1999), who claimed that the interannual variance of the Aleutian low was larger during the positive phases of the PDO but smaller during the negative phases of the PDO. This suggests that, although the SST in the North Pacific is relatively independent of the North Atlantic climate system, during the positive phases of the PDO, the large interannual variance in the

**FIG. 8.** Same as in Fig. 6, but for ALI.
North Pacific winter SLP may set up a link in the atmosphere through planetary wave activity from the North Pacific toward the Arctic and the North Atlantic, which may be triggered by anomalous SLPs associated with the Aleutian low (Minobe and Mantua 1999; Honda et al. 2001).

The temporal variability in the correlation between the SLP field over the North Pacific and North Atlantic Oceans may also explain the results obtained by Zhao and Moore (2004), who found that the nonlinear expression of the NAO in the SLP field over the North Pacific was a function of the amplitude but not sign of the NAO. Zhao and Moore (2004) used the entire NCEP reanalysis time period, 1948–2000, in their analysis and therefore included a regime shift of the PDO. Other studies have also noted a nonlinearity in the response of the climate system to the PDO. For example, Gershunov et al. (2001) found that the response to ENSO over the United States was a function of the phase of the PDO. Hoerling et al. (1997, 2001) have also identified a nonlinear relationship between sea surface temperature and deep convection over the tropical Pacific that was responsible for the approximate 35° longitudinal shift in the centers of action of the PNA between warm and

**Fig. 9.** Same as in Fig. 6, but for PDOI.
cold phases of ENSO. The subperiods identified in this study are consistent with those in Kutzbach (1970), who identified a change in the EOF1 of the wintertime SLP field around the mid-1920s and the mid-1950s. An advantage of using the PDOI as opposed to the ALI as the index for North Pacific climate variability is the existence of paleoclimate reconstructions that, along with similar reconstructions for the NAOI, allow one to consider behavior in the preinstrumental period, that is, before the 1920s when SLP observations in the North Pacific are sparse. With these reconstructions, we have identified one additional reversal in the sign of the correlation between the NAOI and the PDOI from 1870 to 1910 that corresponds to a time when the reconstructed PDOI was undergoing a change in sign (D’Arrigo et al. 2001). In addition, we have identified a change in the nature of the correlation between the reconstructed PDOI and NAOI around 1850. Prior to this time, there was no evidence of the multidecadal reversals in the sign of the correlation that was present after 1850.

This suggests that there might be a fundamental change in the variability of the winter NH atmospheric circulation across 1850, nominally considered to be the end of the Little Ice Age (Denton and Karlen 1973; Bond et al. 1999). In this regard, it is interesting to note that D’Arrigo et al. (2001) noted a change in the frequency characteristics of their reconstructed PDOI around 1850 with a transition from decadal to interannual power occurring around this time. This enhanced interannual variability in the North Pacific may have contributed the interdecadal linkage between the two oceanic basins. The annual snow accumulation time series from the high elevation ice cores from Mount Logan in the Gulf of Alaska region and the southern Himalayas also underwent a transition around 1850 from a regime with no trend to one in which snow accumulation began to increase, in the case of Mount Logan, or decrease, in the case of the southern Himalayas. This behavior was postulated to result from an intensification of planetary wave activity over the North Pacific and North America (Moore et al. 2002) and a weakening of the Walker circulation that occurred around 1850 (Moore et al. 2005; Zhao and Moore 2006). Other studies have indicated that during the Little Ice Age the trade winds across the tropical Pacific were stronger than today and that the middle of the nineteenth century was the period when they weakened,

FIG. 10. The 21-yr MWC coefficients between the reconstructed NAOI time series and the reconstructed PDOI (solid). The thick dashed curve is the same MWC except for the instrumental indexes. The horizontal dashed lines are the 90% confident level using the resampling test. Please note that the PDOI was inverted before calculating the MWC so as to facilitate comparison with Fig. 3.
attaining more modern values (Hendy et al. 2002; Oppo et al. 2003).

Acknowledgments. The authors wish to thank the anonymous reviewers for their insightful critiques that led to significant changes and improvements of the manuscript. The Natural Sciences and Engineering Research Council of Canada and the Canadian Foundation for Climate and Atmospheric Sciences through the Polar Climate Stability Network funded this research. The NAO index was provided by the Climate Research Unit of the University of East Anglia. The ALI was provided by the Climate Analysis Section, NCAR, Boulder, Colorado. The PDOI and AOI were downloaded from the JISAO website.

REFERENCES


