A Well-Verified, Multiproxy Reconstruction of the Winter North Atlantic Oscillation Index since A.D. 1400*

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ABSTRACT

A new, well-verified, multiproxy reconstruction of the winter North Atlantic Oscillation (NAO) index is described that can be used to examine the variability of the NAO prior to twentieth century greenhouse forcing. It covers the period A.D. 1400–1979 and successfully verifies against independent estimates of the winter NAO index from European instrumental and noninstrumental data as far back as 1500. The best validation occurs at interannual timescales and the weakest at multidecadal periods. This result is a significant improvement over previous proxy-based estimates, which often failed to verify prior to 1850, and is related to the use of an extended reconstruction model calibration period that reduced an apparent bias in selected proxies associated with the impact of anomalous twentieth century winter NAO variability on climate teleconnections over North Atlantic sector land areas. Although twentieth century NAO variability is somewhat unusual, comparable periods of persistent positive-phase NAO are reconstructed to have occurred in the past, especially before 1650.

1. Introduction

The North Atlantic Oscillation (NAO) is a key pattern of internal climate variability in the extratropical Northern Hemisphere winter, influencing temperature, precipitation, and atmospheric circulation over a wide region (Hurrell 1995; Hurrell and van Loon 1997). The NAO impacts Arctic sea ice export (Hilmer and Jung 2000), Tigris–Euphrates streamflow (Cullen and deMenocal 2000), plant and animal populations (D'Arrigo et al. 1993; Post and Stenseth 1999), and human activities both present and past (Cullen and deMenocal 2002; Weiss 2002). It has also been linked to widespread warming over the past few decades (Hurrell 1996) and climate anomalies during the ''Little Ice Age'' (Luterbacher et al. 1999, 2001; Shindell et al. 2001). Therefore, it is understandable that a great deal of effort is being spent on observing and modeling the dynamics of the NAO in order to develop a predictive model of its behavior. At the same time, there is a ''pressing need'' to develop longer records of the NAO to place recent variability in a better long-term context (Jones et al. 2001).

The NAO is defined by a seesaw of atmospheric mass between its northern pole near Iceland (the Icelandic low) and its southern pole near the Azores (the Azores high). Its strongest and most climatologically effective expression occurs during the cold season months (Rogers 1984; Hurrell 1995; Jones et al. 1997) when the north–south pressure gradient is greatest between its quasi-stationary centers of action. As such, a convenient expression of NAO behavior over time is usually obtained by calculating the normalized sea level pressure (SLP) difference between the Azores high and Icelandic low for the 4-month winter (December–March) season (Hurrell 1995). This index is calculated using long instrumental SLP records from Iceland and either Ponta Delgadas in the Azores (Rogers 1984), Lisbon in Portugal (Hurrell 1995), or Gibraltar in Spain (Jones et al. 1997). The resulting winter NAO index is little affected by the choice of the southern SLP record.

The power spectrum of the annual winter NAO index indicates the presence of significant band-limited power at 7–25-yr timescales (Rogers 1984; Cook et al. 1998; Mann 2002). Schneider and Schonwiese (1989) also report significant spectral power at 1.7 and 2.2 yr in monthly NAO data. In addition, a multidecadal (50–70 yr) climate signal centered in the North Atlantic has been identified in instrumental records (Schlesinger and

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Ramankutty 1994; Mann and Park 1994) and proxy data (Delworth and Mann 2000). This signal projects only weakly onto the NAO and appears to be governed by distinct oceanic dynamics (Delworth et al. 1997), but during the twentieth century, when this signal is particularly strong, it might contaminate the true underlying NAO signal. Cook et al. (1998), for example, also noted the existence of a 70-yr spectral peak in their NAO reconstruction, but only when the twentieth century data were included. However, a signal of this duration is difficult to characterize and test when its period approaches the length of climate record being studied. Consequently, much longer records are needed to establish the low-frequency variability in the NAO. Using early SLP records from Iceland and Gibraltar, Jones et al. (1997) was able to extend the instrumental NAO index record from 1874 (Rogers 1984) and 1864 (Hurrell 1995) back to 1821. These additional years of data are extremely valuable, but still insufficient to rigorously evaluate the multidecadal behavior of the NAO as suggested by the Schlesinger and Ramankutty (1994) results. They also do not provide an adequate ''natural'' baseline of NAO activity prior to the twentieth century impact of greenhouse gases on climate (e.g., Paeth et al. 1999; Ulbrich and Christoph 1999).

To this end, well-dated and highly resolved proxy records of North Atlantic climate variability have been used to reconstruct the winter NAO index. These include annual tree-ring chronologies from circum-North Atlantic land areas (Cook et al. 1998), an ice core accumulation record from Greenland (Appenzeller et al. 1998), and multiproxy assemblages (Stockton and Glueck 1999; Cullen et al. 2001; Glueck and Stockton 2001; Mann 2002). Most of these proxy-based reconstructions begin in 1700 or earlier. Alternately, early European instrumental SLP, temperature, and precipitation records have been used by Luterbacher et al. (1999) to reconstruct the winter NAO index back to 1675, with more recent extensions back to 1500 (Luterbacher et al. 2001). Each of these NAO index reconstructions was subjected to rigorous statistical calibration/verification procedures and all appeared to be valid. However, the validity of the proxy-based reconstructions was called into question by Schmutz et al. (2000) when three of them (Cook et al. 1998; Appenzeller et al. 1998; Stockton and Glueck 1999) failed to verify against the NAO index of Jones et al. (1997) prior to 1850. In contrast, a modified version of the Luterbacher et al. (1999) reconstruction (minus Gibraltar and Iceland SLP data) correlated significantly with the Jones et al. (1997) series back to 1821. Further comparisons of the proxy-based reconstructions with the longer Luterbacher et al. (1999) reconstruction indicated that their failure to verify extended back into the eighteenth century as well. These results led Schmutz et al. (2000) to conclude that the proxy-based NAO index reconstructions could not be trusted. A comprehensive review of these results by Cook (2002), which included testing some new proxy-based NAO index reconstructions described in Cullen et al. (2001), came to the same basic conclusion. The proxy-based estimates could not be trusted prior to 1850.

Here, we describe a new multiproxy reconstruction of the winter NAO index back to 1400 that verifies significantly against all extended instrumental-based NAO indices produced thus far. This improvement arises from the use of an expanded network of proxies from circum-North Atlantic land areas and an extended calibration period that is less affected by possible twentieth century bias in the teleconnection between the NAO and climate over that region. Proxy-based reconstructions of the kind described here all rely on the connection between the NAO and climate (e.g., temperature and precipitation) over the areas where the proxies originate. As such, the proxies are recording the influence or impact of the NAO on climate somewhat removed from its actual SLP centers of action. This distinction must be kept in mind because no proxies directly record SLP information. While the winter NAO index reconstruction presented here undoubtedly contains useful information on the behavior of the NAO itself, it also contains information associated with the teleconnected impacts of the NAO on climate as well.

2. Long instrumental winter NAO index records

Three expressions of the winter NAO index, all based on European instrumental climate records, are used here for statistical calibration of the proxies and verification of their estimates. Figure 1a is the Gibraltar–Iceland index estimated directly from monthly station SLP data (Jones et al. 1997, referred to hereafter as J97). It begins in 1826 when the winter season becomes serially complete. Figure 1b shows estimates of the same winter NAO index back to 1781 using the SLP reconstructions for grid points over Iceland and Gibraltar (Jones et al. 1999, hereafter J99). These data were kindly provided by P. D. Jones. Figure 1c is a recently updated and revised winter NAO index reconstruction (Luterbacher et al. 2001, hereafter L01) extending back to 1659 based on early European instrumental SLP, temperature, and precipitation records recovered as part of the European project ADVICE (Annual to Decadal Variability in Climate in Europe). This series was kindly provided by J. Luterbacher.

It is apparent that the three NAO index series in Fig. 1 are extremely similar over the 1826–1990 common period, in part because they share some common data. The J97 and J99 series have a correlation of 0.96, and the correlations of L01 with J97 and J99 are 0.89 and 0.87, respectively. The correlation between L01 and J99 is also 0.87 over the earlier 1781–1825 overlap period. Besides sharing some common data, the three NAO index series are similar because they share the same European climate system that is affected by the NAO. This fact suggests that they may unduly emphasize climate

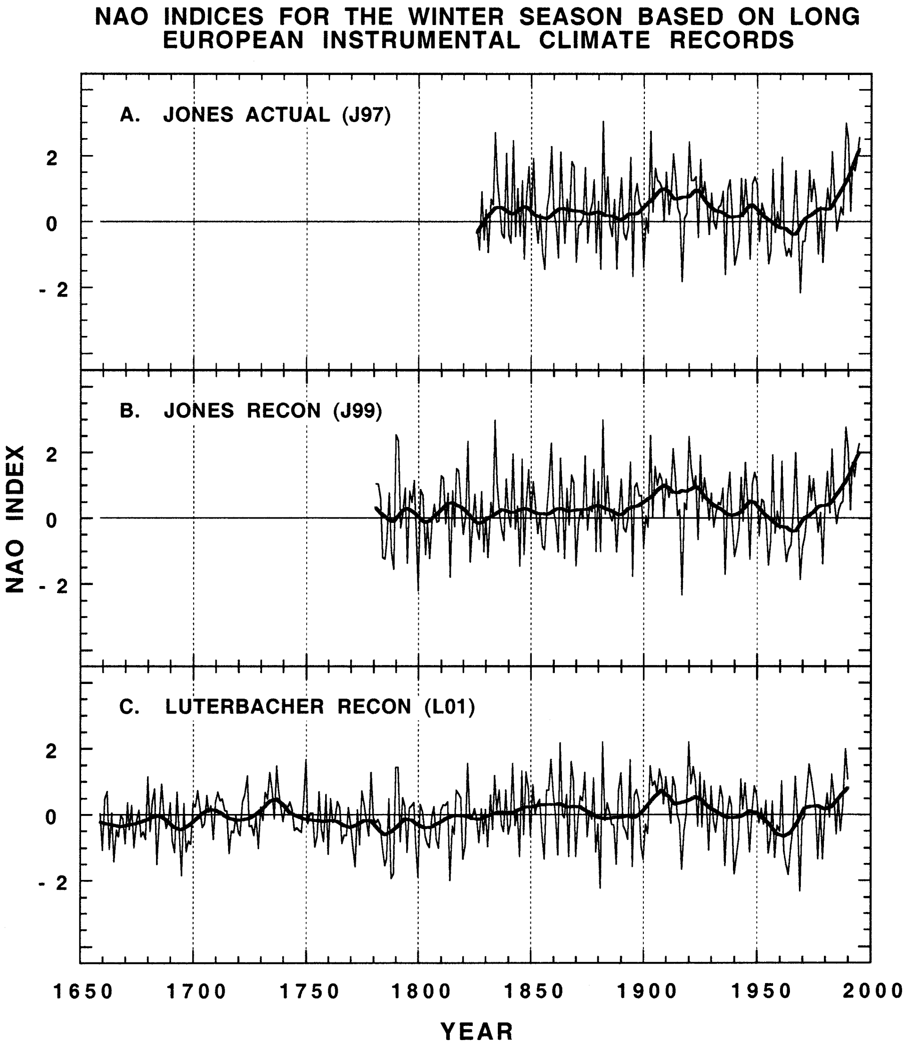

FIG. 1. Long instrumental-based estimates of the winter NAO index from European data. The two Jones series are based strictly on SLP data, while the Luterbacher series is based on SLP, temperature, and precipitation data.

impacts over Europe at the expense of those over eastern North America where the NAO also has an influence on climate (Hurrell and van Loon 1997).

It is also apparent that the behavior of the NAO during the twentieth century has been anomalously strong, with persistent periods of positive and negative phases of the NAO. This may be an expression of both the multidecadal variability described earlier, and also a manifestation of greenhouse gas forcing (Paeth et al. 1999; Ulbrich and Christoph 1999; Shindell et al. 1999) and increasing tropical sea surface temperatures (Hoerling et al. 2001). There is no evidence of a persistent 50–70 yr signal prior to 1900 (e.g., in J97, J99, and L01), which suggests that it is not a long-term feature of the NAO. Thus, any calibrations between the NAO index and selected proxies may be significantly biased if the calibration interval is dominated by twentieth century data. Such would be the case using either the Rogers (1984) index back to 1874 or the Hurrell (1995) index back to 1864. Therefore, it is probably not coincidental that the Cook et al. (1998), Appenzeller et al. (1998), and Stockton and Glueck (1999) reconstructions, calibrated using the Rogers and Hurrell records, failed in the tests performed by Schmutz et al. (2000) using longer instrumental-based NAO indices.

Given the possible calibration bias associated with anomalous twentieth century NAO variability, expanding the calibration time period to include more of the nineteenth century is desirable. This is possible now because of the long instrumental-based NAO indices shown in Fig. 1. Specifically, we will use the J99 series over the 1826–1974 interval to calibrate the proxies, which have a common end year of 1974. This period covers the serially complete interval of J97 and gives nearly equal weighting to both the nineteenth and twentieth centuries. The 1781–1825 data in J99 and the

1659–1825 data from L01 will be withheld from all calibration modeling for verifying the accuracy of the proxy estimates.

3. The proxy data and regression procedure

The teleconnection studies of Hurrell and van Loon (1997) indicate that the NAO significantly impacts climate over parts of eastern North America, Europe, North Africa, and the eastern Mediterranean region. Fortunately, these are regions with abundant supplies of centuries-long, climatically sensitive, annual tree-ring chronologies, some of which have already been used in previous reconstructions of the winter NAO index (Cook et al. 1998; Stockton and Glueck 1999; Cullen et al. 2001; Glueck and Stockton 2001; Mann 2002). In addition, a number of well-dated and highly resolved ice core records from Greenland (both stable isotope ratios and accumulation rates), covering the past several hundred years, are available for reconstructing the NAO index. As before, some of them have been used previously for that purpose (Appenzeller et al. 1998; Stockton and Glueck 1999; Glueck and Stockton 2001). Most of these records are in the public domain and can be downloaded anonymously from the National Oceanic and Atmospheric Administration (NOAA) National Geophysical Data Center, World Data Center for Paleoclimatology web site (http://www.ngdc.noaa.gov/paleo/data.html). However, a number of tree-ring series used here, especially from Europe, are not yet in the public domain.

A total of 367 candidate predictors of the winter NAO index were amassed, all covering the common interval 1750–1974. Of those, 329 extend back to 1700, 141 back to 1600, 86 back to 1500, and 49 back to 1400. In order to take full advantage of the longer proxies, a series of nested principal component regression (PCR) models were developed in which the common starting year of the candidate predictors was stepped backward in 25-yr increments. This resulted in a total of 15 regression model runs, each with its own calibration and verification statistics.

The PCR procedure used here closely follows that described in Cook et al. (1999) for reconstructing drought indices from tree rings [for other examples of PCR in dendroclimatology, see Fritts (1976), Briffa et al. (1986), and Cook et al. (1994)]. Here we give a brief generic description of the method. We begin with the classic multiple linear regression model

$$\mathbf{y} = \mathbf{UB} + \mathbf{e}, \tag{1}$$

where **y** is the vector of standardized (i.e., zero mean, unit standard deviation) dependent or predictand data (in this case NAO indices), **U** is the matrix of orthogonal principal components (PCs) used for prediction (the proxies), **B** is the matrix of standardized regression coefficients (beta weights), and **e** is the vector of regression model errors. The regression model is developed over the *calibration* time period common to the predictors and predictand (here 1826–1974), with the pre-1826 instrumental NAO data reserved for regression model *verification* tests of the tree-ring model estimates. Because the PCs in the calibration period are orthogonal, the beta weights in **B** are nothing more than the Pearson simple correlations between the predictors and predictand.

The actual proxies used as predictors are related to their PCs as

$$\mathbf{U} = \mathbf{XF}, \tag{2}$$

where **X** is the matrix of standardized tree-ring chronologies used as predictors and **F** is the orthonormal matrix of column eigenvectors calculated from the correlation matrix of **X**. Once the regression coefficients **B** have been estimated for the calibration period, they are applied to the PCs in **U** for as far back as the proxies go (say 1750) to produce a series of estimates

$$\hat{\mathbf{y}} = \mathbf{UB}. \tag{3}$$

The resulting standardized estimates are than back-transformed to original units using the mean and standard deviation of the predictand data, thus producing the climate reconstruction from the proxies.

The PCR approach is similar to that used by Luterbacher et al. (2001) to reconstruct the winter NAO index using early European instrumental records of varying lengths. However, unlike the European climate data used by Luterbacher et al. (2001), it is difficult to know a priori which proxies are potentially useful records of climate associated with winter NAO variability. Therefore, some sort of predictor variable screening is desirable. As described in Cook et al. (1999), this is done on the individual proxies before principal components analysis is applied to those that are retained. In this case, only those proxies that correlated significantly ($r > |0.15|$; $p < 0.10$) with the winter NAO index over the 1826–1974 calibration period were retained.

Screening at the level just described resulted in a dramatic reduction in the number of proxies actually entered into each nested PCR run, typically about one-third of the original number of candidate predictors. This procedure reduces the noise level in the resulting PCs associated with the predictand by effectively prefiltering the proxy variable set to eliminate useless data. The dimensionality of the candidate predictors was further reduced within PCR itself by only retaining those PCs having eigenvalues >1.0 (the Kaiser–Guttman rule). This rule reduced the number of retained PCs to about one-third again on average. Finally, the best subset of the retained PCs entered into the actual regression model was determined using the minimum Akaike information criterion (Akaike 1974), with a small-sample bias correction applied (Hurvich and Tsai 1989).

An apparent drawback of PCR relates to the fact that the regression coefficients in (1) are on the PCs themselves, not on the original variables. This makes the interpretation of the PCR model difficult. However, it is straightforward to transform the beta weights from PC space to real data space as

$$\boldsymbol{\beta} = \mathbf{FB}, \quad (4)$$

where $\boldsymbol{\beta}$ is the vector of beta weights on the original proxies, $\mathbf{F}$ is the matrix of retained eigenvectors, and $\mathbf{B}$ is the vector of beta weights in PC space. See Cook et al. (1994) for details.

The calibration and verification statistics described here are the same as those provided by Luterbacher et al. (1999) for their reconstruction: the coefficient of multiple determination (R^2) for the calibration period, and the square of the Pearson product-moment correlation (R*R) and reduction of error (RE) for the verification period. These statistics are all measures of variance in common between the actual (J99) and estimated (multiproxy) winter NAO indices. See Cook et al. (1999) for more details on these statistics and their comparative differences.

Figure 2 shows the map of proxies retained after the initial screening for correlation with the J99 winter NAO index (Fig. 2a) and the beta weights (Fig. 2b) of those series used in reconstructing the winter NAO index. All series cover the common interval 1750–1974. However, some of the useful proxies extend out to 1979, which allowed the reconstruction to be extended forward to that year. The distribution of sites is well balanced on both sides of the Atlantic and agrees well with the NAO teleconnection patterns with climate found by Hurrell (1995) and Hurrell and van Loon (1997) over the North Atlantic sector. This result suggests the additional data from North America may provide a more complete estimate of winter NAO variability than that available from European-only data. The beta weights support this suggestion. The relative explanatory power of the proxies is evenly distributed across the network.

4. The new winter NAO index reconstruction

Figure 3 shows the winter NAO index reconstruction (Fig. 3a), with the number of proxies available for each 25-yr step back in time from 1750, using the network of proxies shown in Fig. 2. For comparing the reconstruction to the most recent changes in winter NAO, the series has also been extended from 1979 to 2001 using instrumental NAO data obtained from the Climatic Research Unit, University of East Anglia (http://www.cru.uea.ac.uk/cru/data/nao.htm). The variance of the data used for the extension has been reduced to reflect the lost variance due to regression.

A comparison of twentieth century NAO behavior with that estimated back to 1400 indicates that NAO variability over the most recent 100 years is unusual, but not unique. This finding is consistent with that of Glueck and Stockton (2001) and Luterbacher et al. (2001). In particular, the fifteenth and sixteenth centuries appear to have experienced episodes of persistent positive-phase NAO behavior that are comparable to those seen in the twentieth century. In contrast, during the 1640–1880 interval the NAO appears to have been less vigorous. This may be an expression of the impact of Little Ice Age cooling on climate in the North Atlantic sector.

Below Fig. 3a are plots of the calibration and verification statistics described earlier (Fig. 3b). The calibration R^2 ranges from 0.636 in the post-1700 period to 0.305 for the 1400–1424 period. This decline in R^2 closely tracks the decline in the number of proxies available in each time period, a result that is consistent with expectation (Rencher and Pun 1980). In contrast, the verification R*R and RE statistics remain remarkably constant over time, with ranges of 0.269–0.367 and 0.237–0.350, respectively. With 45 paired observations in the 1781–1825 verification period to compare, these results strongly support the validity of this multiproxy winter NAO index reconstruction.

To further test its validity, the reconstruction was compared to the modified and updated L01 reconstruction over the 1659–1979 period common to each series using the Kalman filter as a dynamic regression modeling procedure (Harvey 1981; Visser and Molenaar 1988; Van Deusen 1990). Unlike the ad hoc running-correlation procedure used by Schmutz et al. (2000) to compare estimates of the NAO index over time, the Kalman filter method allows for the identification of time-dependence between predictor and predictand variables in a totally objective way using maximum likelihood estimation (MLE). This is accomplished by casting the simple regression model into state-space form with a state equation

$$Y_t = a_t X_t + e_t \quad (5)$$

and a state transition equation

$$a_t = a_{t-1} + n_t. \quad (6)$$

The disturbances e_t and n_t are both scalar in the simple regression case and are assumed to be normally and independently distributed as NID(0, σ^2) and NID(0, $\sigma^2 q_t$), respectively. In this form, the regression coefficient a_t is allowed to vary as a random walk, with its variance dependent on n_t. If var(n_t) = 0 (i.e., $q_t = 0$), the model reduces to a classical simple regression model with a constant coefficient a. For var(n_t) > 0 (i.e., $q_t > 0$), the simple regression model becomes dynamic and locally adapts through a_t to the changing relationship between the predictor X and predictand Y. Determining if var(n_t) > 0 is accomplished by MLE using as the initial state var(n_t) = 0. The MLE solution also provides standard errors for determining the significance of a_t over time. See Harvey (1981) and Visser and Molenaar (1988) for details.

The results of the Kalman filter comparison are shown in Fig. 4 in the form of a_t plotted as a function of time with its two standard error limits. As long as the error limits do not cross zero, a_t is considered to be signifi-

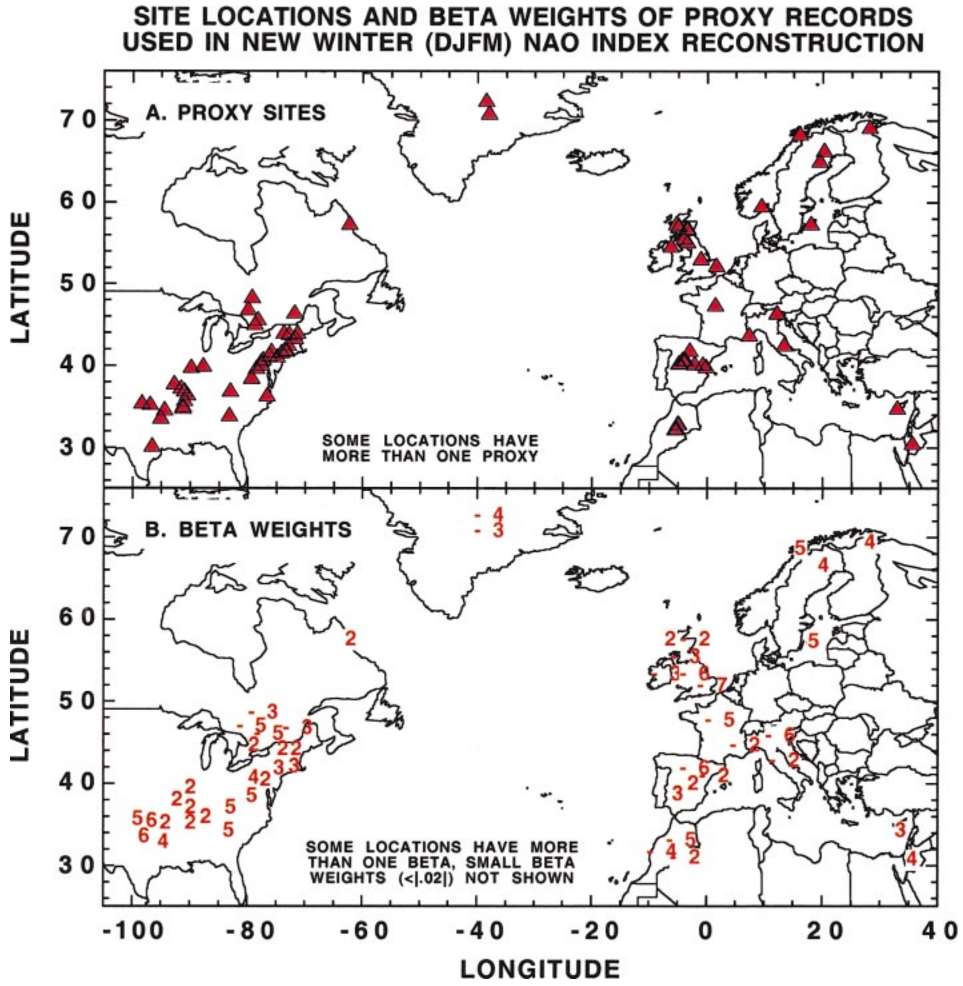

FIG. 2. Maps showing the distribution of (a) proxies used to reconstruct the winter NAO index and (b) their standardized regression coefficients or beta weights. The beta weights show the relative explanatory power of the proxies, which is seen to be homogeneously distributed across the network.

cantly different from zero. The MLE solution indicates that the relationship between the multiproxy reconstruction and L01 is slightly time dependent, but always positive and statistically significant at the 2 standard error limit. Therefore, this new NAO reconstruction verifies against the updated and revised L01 reconstruction back to 1659, a result that is clearly superior to the negative findings of Schmutz et al. (2000). Unfortunately, the instrumental-based reconstruction of Luterbacher et al. (1999) used by Schmutz et al. (2000) is significantly different from the better quality one of Luterbacher et al. (2001) used here. Therefore, it is not possible to know how the comparisons made by Schmutz et al. (2000) would change using the L01 record.

Also shown in Fig. 4 are four correlations that reveal more discretely how the actual correlation between the two reconstructions has changed. These correlations are shown for the 1659–1780 ($r = 0.53$), 1781–1825 ($r = 0.52$), 1826–1979 ($r = 0.68$), and 1675–1780 (minus 1714–1721) ($r = 0.59$) time periods. The 1826–1979 period is effectively the same as the calibration period used here and is provided for comparison with the R^2 results shown in Fig. 3. The 1781–1825 period is the same as the verification period used previously and its r agrees favorably (when squared) with that shown in Fig. 3. For the 1659–1780, which is independent of the previous verification tests, the verification r remains about the same. However, this result may underestimate the true fidelity of the multiproxy reconstruction. Only the 1675–1713 and 1722–1780 periods of L01 used early SLP data, which significantly improved the calibration/verification statistics (see Fig. 2 in Luterbacher et al. 2001) over other periods for which SLP data were not available. When only those years of reconstructed NAO index are compared, the verification r increases to 0.59. In contrast, when the reconstructions for the 1659–1674 and 1714–1721 periods based on no SLP data are correlated, the verification r drops to 0.36. Thus, this drop appears to come primarily from a weakness

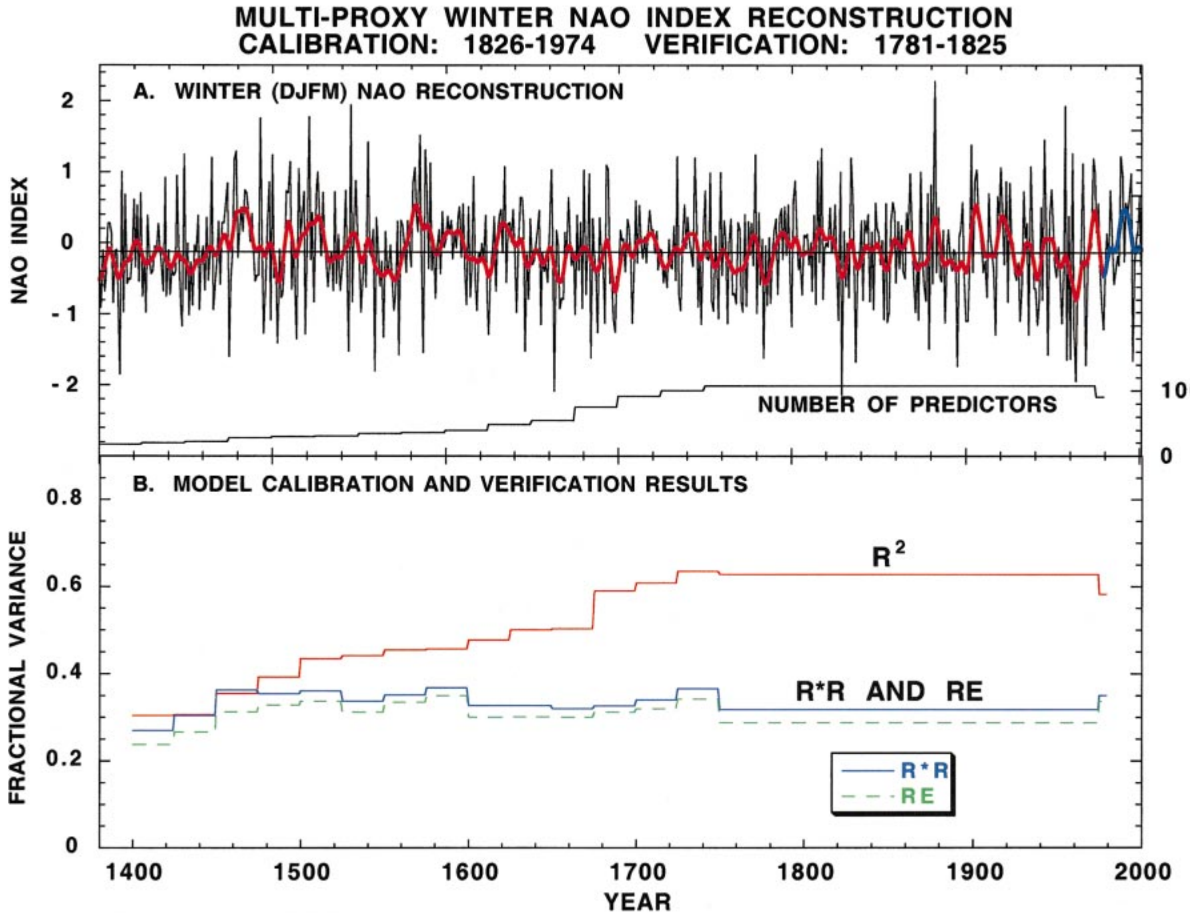

FIG. 3. (a) The multiproxy reconstruction of the winter NAO index and the number of proxies used, with (b) the time-varying calibration and verification statistics over the period 1400–1979. The reconstruction has been extended from 1979 to 2001 using appropriately scaled instrumental NAO data for comparison with variability during prior centuries.

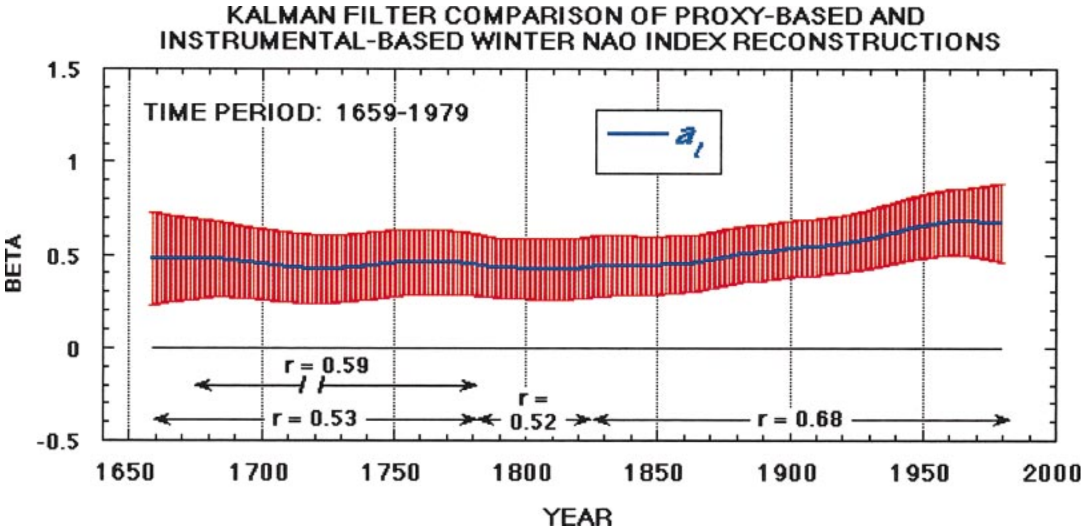

FIG. 4. Comparison of the multiproxy NAO index reconstruction with L01 (see Fig. 1c) using the Kalman filter as a dynamic regression modeling procedure. The relationship is time dependent but always statistically significant back to 1659. The simple correlations show more discretely how the relationship changes.

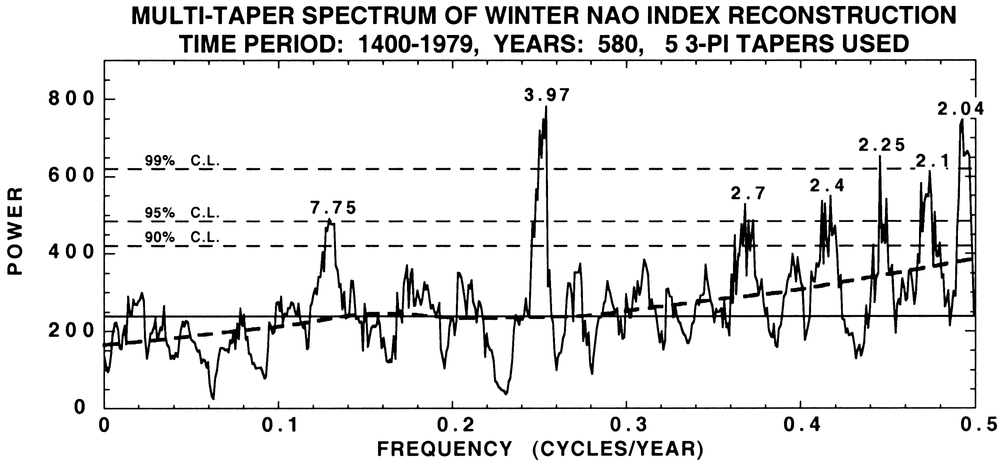

FIG. 5. The multitaper power spectrum of the new winter NAO index reconstruction. The reconstruction is dominated by band-limited power at periods shorter than 10 yr and behaves overall as a ''blue noise'' process with power increasing with frequency (thick dashed line). The latter result is consistent with the spectrum of the instrumental SLP-based J99 record in Fig. 1b.

in the L01 reconstruction when it is not based on SLP data. Therefore, we regard the other verification results as being more indicative of the true accuracy of the multiproxy reconstruction.

One additional comparison was made between the new multiproxy reconstruction and estimates of the winter NAO index produced by Luterbacher et al. (2001) for the period 1500–1658 from early noninstrumental European data. The r between these two series is 0.22, which is much lower than before, but still significant well above the 95% confidence level. So, assuming that these European estimates of the winter NAO index are reasonably accurate back to 1500, the multiproxy reconstruction can be regarded as being significantly verified against independently derived estimates covering the past 479 yr. This result is quite remarkable given the earlier negative results of Schmutz et al. (2000) and Cook (2002) in testing some previous proxy-based NAO index reconstructions.

Given the successful long-term verification of this reconstruction, we computed its power spectrum to determine the degree to which it has maintained some of the same band-limited power found in the shorter instrumental records (e.g., Rogers 1984; Cook et al. 1998; Mann 2002). The spectrum (Fig. 5) shows that most of the significant band-limited variance is restricted to periods of less then 10 yr, with particular concentrations around 7.7 and 4 yr and in the quasi-biennial (i.e., 2–3 yr) band. This is consistent with previous results, which suggests that these band-limited properties of the winter NAO are organized long-term features of North Atlantic SLP. There is no indication of a distinct multidecadal oscillation in the spectrum, although there is a ''lump'' of variance with a mean period of about 50–60 yr. However, it is important to note that there is a modest decrease in calibrated variance at periods greater than about 10 yr, and even more significant loss of variance at timescales greater than 50 yr (as estimated from the spectral coherence between the J99 and reconstructed series), so definitive conclusions cannot be reached regarding the multidecadal and century-scale variations in the NAO at this stage. The cause of this decline in multidecadal fidelity is unclear. A detailed examination of the proxies used in the reconstruction, and how they were developed, will be necessary to diagnose this deficiency.

Finally, unlike many climatological time series that behave as ''red noise'' processes, the winter NAO index reconstruction behaves more like a ''blue noise'' process. That is, most of its variance is concentrated in the high-frequency end of the spectrum. The lowest robust smoothing of the spectrum (the thick dashed line in Fig. 5) shows this overall increasing trend in variance. This property is unlikely to be an artifact of the statistical processing applied to the proxies because the spectrum of the J99 instrumental series (not shown) behaves in much the same way. Therefore, the ''blueness'' of the reconstruction appears to be a real feature of SLP in the North Atlantic as measured by the winter NAO index. Even so, the interpretation of the lowest-frequency variability, as discussed before, is limited by the possible loss of calibrated variance at very low frequencies.

5. Why did previous proxy-based reconstructions fail?

The successful reconstruction of the winter NAO index begs the question: why did the previous reconstructions fail to verify against the instrumental-based NAO indices prior to 1850? The use here of a large number of widely distributed proxies may have contributed to more robust estimates of the NAO index back in time. Yet, the verification statistics in Fig. 3 remain very stable even with large reductions in the number of proxies

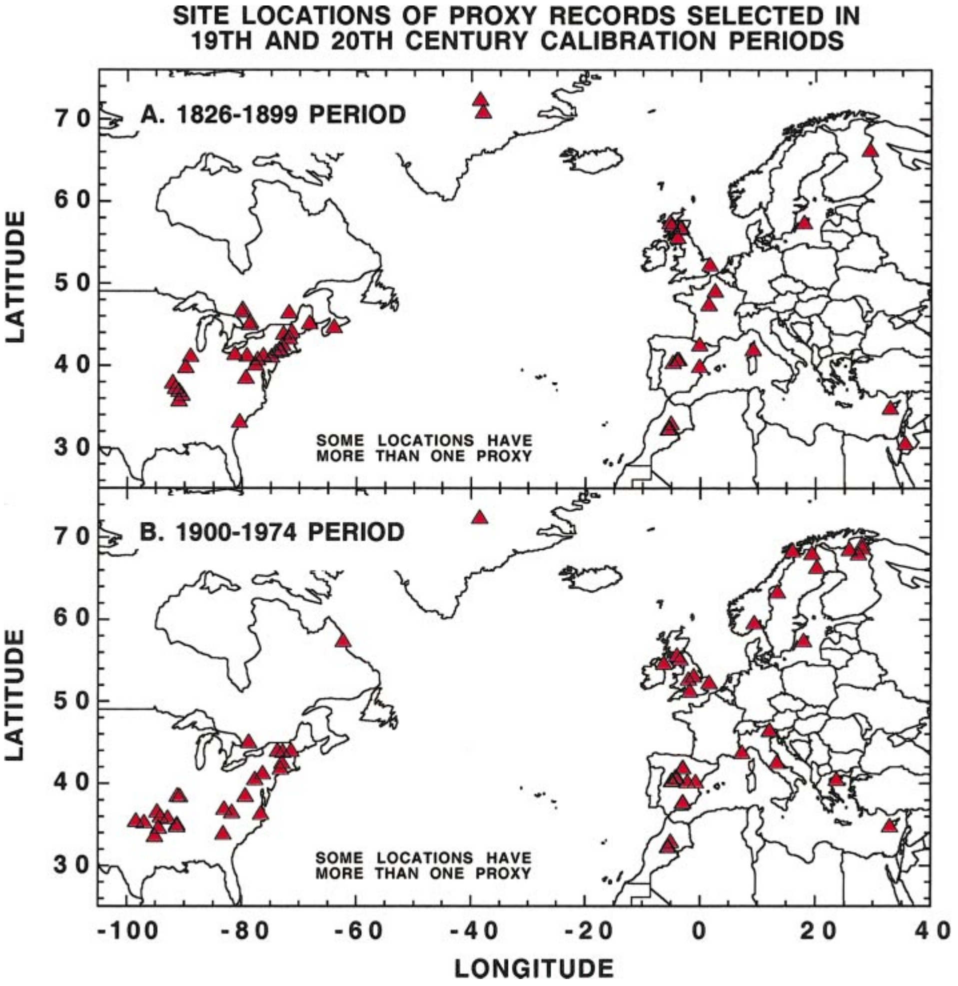

FIG. 6. Maps of selected proxies for two equal calibration periods in the nineteenth and twentieth centuries. In each case, the selected proxies are significantly correlated ($p < 0.05$) with the J99 index. The change in the distribution of proxies, especially in Fennoscandia and the lower Mississippi Valley, indicates that their selection based mainly on twentieth century data is geographically biased. This finding probably explains why previous proxy-based reconstructions of the NAO index failed to verify prior to 1850 in most cases.

used. Therefore, we believe that the answer lies more likely in the choice of the extended calibration period used here.

As noted earlier, the behavior of the NAO in the twentieth century appears to be more persistent and extreme than that seen in the nineteenth century (see Fig. 1). This need not be a problem in selecting the proxies for reconstruction if the teleconnection pattern between the NAO and circum-North Atlantic climate during the twentieth century (e.g., Hurrell 1995; Hurrell and van Loon 1997) is similar to that in the past. However, we believe that this has not been the case. A possible explanation for anomalous twentieth century behavior in NAO teleconnections is that the NAO (or the related Arctic Oscillation or ''AO''; see e.g., Thompson and Wallace 2000; Thompson et al. 2000) has been overprinted in recent decades by anthropogenic forcing. The anthropogenic signal may consist of both an NAO/AO component (Shindell et al. 1999; Paeth et al. 1999; Hoerling et al. 2001) and a distinct, more complex pattern of anthropogenic sea level pressure change. The latter pattern projects a spurious apparent NAO signature (e.g., Broccoli et al. 1998). If this were the case, the anthropogenic pattern would effectively ''contaminate'' the twentieth century with a related but distinct pattern of variability, and would complicate the accurate calibration of the NAO component in a restricted twentieth century interval. This interpretation is consistent with recent work by Rutherford et al. (2001, submitted to *J. Climate*) using both control and forced integrations of the GFDL R30 coupled model. This latter work shows that while there is no evident bias in reconstructing past large-scale climate patterns from sparse data based on calibration during a nonstationary (anthropogenic forced) interval that is long enough to resolve both the forced and internal patterns of variability, a biased reconstruction of the past *will* result if that interval is too restricted in duration. This could explain the increase in the fidelity of the NAO reconstruction based on a nineteenth-to-twentieth century calibration relative to that based on mostly twentieth century data.

To illustrate the changes in the apparent teleconnections of the NAO during the nineteenth and twentieth century, we show in Fig. 6 maps of the proxies selected by the same screening procedure described above for two different calibration periods: 1826–99 and 1900–74. This split breaks the J97 series into equal length subseries that also have distinctly different low-frequency characteristics (see Fig. 1a). A comparison of the maps indicates that the 1900–74 calibration period chooses many more proxies in the Fennoscandia and lower Mississippi Valley regions than does the 1824–99 calibration period. This change is unlikely to be related to possible changing quality of the proxies themselves. The tree-ring chronologies are all highly replicated in the time periods being tested. Rather, we believe that the change in the geographical distribution of selected proxies is evidence for a change in the teleconnection of the winter NAO with circum-North Atlantic climate. Consequently, the proxies used in previous reconstructions [e.g., those tested by Schmutz et al. (2000)] may have been geographically biased toward regions that were strongly affected by the anomalous twentieth century behavior of the NAO. Prior to the twentieth century, and especially before 1850, this bias led to a significant loss of reconstruction skill. By expanding the calibration period to equally weight the nineteenth and twentieth centuries, it was possible to identify a less geographically biased set of proxies (cf. Figs. 2 and 6) to reconstruct the winter NAO index. Other factors may be playing lesser roles, but we believe that this is probably the first-order cause of the extended verification success reported here.

6. Conclusions

We have presented here the first multiproxy reconstruction of the winter NAO index that verifies well against previous estimates of this index based on early European instrumental (Jones et al. 1999; Luterbacher et al. 1999) and noninstrumental (Luterbacher et al. 2001) data. The verification is statistically significant ($p < 0.05$) back to 1500, making our reconstruction one of the most highly validated proxy-based records of past climate yet produced. Thus, we are confident that this record contains useful information on winter NAO variability at least back to 1500 and probably back to its beginning in 1400.

The reconstruction's improved verification is most likely due to the use of an extended calibration period that reduced an apparent geographic bias in the selection of proxies used for reconstruction in previous studies. This bias appears to be related to the anomalous nature of NAO teleconnections over circum-North Atlantic land areas during the twentieth century, which may relate to anthropogenically forced nonstationarity of the climate during the twentieth century. Changing teleconnections between ENSO and drought over the United States during the late-nineteenth and twentieth centuries have also been documented by Cole and Cook (1998). So, the possible effect of such changes on the subsequent quality of climate reconstructions must be kept in mind when selecting and calibrating proxies with (mostly) twentieth century climate data using indirect, teleconnected relationships.

The use of this reconstruction for identifying and characterizing long-term teleconnections between the NAO and other climate reconstructions must be done carefully. Because of the extended nature of North Atlantic proxy network used here (Fig. 2), long proxy reconstructions of associated climate variables, like precipitation over Morocco or the lower Mississippi Valley, cannot be compared to our multiproxy reconstruction without the danger of circularity. However, it should be useful for examining potential proxy-based teleconnections in more distant regions of the world potentially affected by the NAO and AO. And, providing that multiproxy reconstructions of other ocean–atmosphere processes like ENSO are based on completely independent data (e.g., Stahle et al. 1998), it will be possible to look for relationships and interactions between these important internal global forcings.

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