

The Role of Boundary Conditions in AMIP-2 Simulations of the NAO

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ABSTRACT

The simulated North Atlantic Oscillation (NAO) teleconnection patterns and their interannual variability are evaluated from a suite of atmospheric models participating in the second phase of the Atmospheric Model Intercomparison Project (AMIP-2). In general the models simulate the observed spatial pattern well, although there are important differences among models. The NAO response to interannual variations in sea surface temperature (SST) and snow-cover boundary forcings are also evaluated. The simulated NAO indices are not correlated with the observed NAO index, despite being forced with observed SSTs, indicating that SSTs are not driving NAO variability in the models. Similarly, although a number of studies have identified a link between Eurasian snow extent and the phase of the NAO, no such link is apparent in the AMIP-2 results. It appears that, within the framework of the AMIP-2 experiments, the NAO is an internal mode of atmospheric variability and that impacts of SSTs and Eurasian snow cover on the phase of the NAO are not discernable. However, these conclusions do not necessarily apply to decadal-scale and longer variability or to coupled atmosphere–ocean models.

1. Introduction

The North Atlantic Oscillation (NAO) is the only Northern Hemisphere (NH) teleconnection pattern that is active during all 12 months of the year, and it is the dominant pattern of interannual variability in winter (Barnston and Livezey 1987). The NAO is characterized by a dipole in the sea level pressure field with one anomaly center over the Arctic, near Greenland, and another center of the opposite sign across the mid-latitude sector of the North Atlantic Ocean. Observational studies have also revealed NAO-like patterns of variability in sea surface temperatures (SSTs), surface air temperatures, and atmospheric geopotential height fields on decadal and longer time scales (e.g., Deser and

Blackmon 1993; Kushnir 1994). The impact of the NAO on surface conditions, particularly over the North Atlantic Ocean, Europe, and western Asia, has been well established (Hurrell and Kushnir 2003). Nevertheless, despite significant research efforts in recent years, many questions remain regarding the influence of boundary forcings on the NAO (Hurrell and Kushnir 2003).

A number of possible NAO boundary forcing mechanisms have been identified. While some portion of the NAO signal is due largely to stochastic variability internal to the atmosphere, boundary forcings are believed to play a potential role in modulating both the phase and amplitude of the NAO, especially at the decadal and longer time scales (Czaja et al. 2003). Numerous modeling studies have linked multiannual and longer time scale NAO variations with North Atlantic SSTs (e.g., Watanabe et al. 1999; Latif et al. 2000; Mehta et al. 2000). Whether the ocean is an important factor in interannual variations of the NAO is of greater debate. While some numerical experiments have succeeded in reproducing interannual NAO variations in atmospheric models either driven by observed

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SSTs or run in a coupled mode (e.g., Rodwell et al. 1999), the result is not universal (Robertson 2001; Josey et al. 2001). Bretherton and Battisti (2000) caution that those model results that reproduce NAO variability when forced by SSTs are largely reproducible by forcing the model with stochastic variations in the atmosphere and are not indicative of an underlying physical process. They argue that the hindcast skill achieved by atmospheric global climate model (AGCM) experiments with prescribed observed SSTs is an artifact of forcing the model with the observed atmospheric low-frequency variability recorded in the observed SSTs. Forcing an ensemble of AGCM simulations with observed SSTs acts as a filter on atmospheric variability and exposes the low-frequency component recorded in the SSTs, which then feeds back on the model atmosphere. Mysak and Venegas (1998) suggest the presence of a feedback loop linking decadal signals in sea ice to the phase of the NAO. At the upper boundary, evidence has been found linking enhanced westerlies in the lower stratosphere, and associated cold polar air temperatures, with positive trends in the phase and strength of the NAO (Thompson et al. 2003).

Several empirical and modeling studies have detected an influence of Eurasian snow-cover extent on the NAO. Cohen and Entekhabi (1999) demonstrate empirically that anomalies in fall Eurasian snow cover are significantly correlated with the winter NAO. While the most significant correlation appears to be with snow cover during the fall season, other studies have confirmed statistically significant correlations with snow cover during other seasons as well (Bamzai 2003; Bojariu and Gimeno 2003; Saito and Cohen 2003; Saunders et al. 2003). One plausible mechanism by which fall Eurasian snow cover might influence the subsequent winter NAO has been suggested by both observational and numerical analyses (Saito et al. 2001; Cohen et al. 2002; Gong et al. 2003). The proposed mechanism involves a series of energy perturbations that propagate through the troposphere and stratosphere. First, diabatic cooling associated with snow-cover anomalies over Siberia perturbs local stationary wave energy forced by the high topography of Asia. For positive (negative) snow anomalies, increased (decreased) upward energy flux perturbs the local troposphere, is propagated into the lower stratosphere, and eventually affects the remote troposphere. The entire cycle, beginning with the precursor regional tropospheric anomaly to the forced anomaly in the stratosphere and eventually to the induced tropospheric hemispheric anomaly, occurs over a time scale of several months. This process is consistent with our knowledge of troposphere-stratosphere coupling (Zhou et al. 2002). However, this mechanism is limited to the cold season.

NAO variations have also been linked with recent climate change. For example, it has been suggested that a significant positive trend in the NAO index during recent decades may be responsible for a large fraction

of the observed warming in the NH (Hurrell 1995; Thompson et al. 2000). Furthermore, modeling studies have been used to link trends in the NAO with various boundary forcings, including SSTs. For example North Atlantic SSTs, have been suggested as being at least partially responsible for forcing the positive trend in the observed NAO (Rodwell et al. 1999). And more recently, Hoerling et al. (2001) present evidence that climate changes in the North Atlantic since the 1950s, including NAO variations, are linked to a progressive warming of the tropical oceans, especially the Indian and Pacific Oceans. As AGCMs are the preferred tool with which to study such climatic variations, it is important to evaluate AGCM performance with regard to interannual NAO variability and its relationship to boundary forcings. In this study, we explore results from AGCM experiments that were performed as part of the second phase of the Atmospheric Model Intercomparison Project (AMIP-2). In particular, we examine the spatial and temporal evolution of the NAO and its relationship to SSTs and snow cover.

2. Data, models, and methods

A variety of methods can be used to calculate an NAO time series index and spatial pattern. For this analysis we use the time series and spatial loading pattern associated with the first empirical orthogonal function (EOF) of the NH winter [December–February (DJF)] sea level pressure (SLP) field obtained from the National Centers for Environmental Prediction (NCEP–National Center for Atmospheric Research (NCAR) Reanalysis (Kalnay et al. 1996) for the years 1979–95, corresponding to the AMIP-2 time domain. Although debate exists as to whether the NAO is a truly independent regional oscillation or rather the regional manifestation of the hemispheric teleconnection pattern known as the Arctic Oscillation (AO; Thompson and Wallace 1998) or more recently as the NH Annular Mode (NAM; Thompson et al. 2003), Hurrell and Kushnir (2003) find that the spatial patterns and time series calculated using the entire NH closely resemble those calculated using only the North Atlantic sector. Station based indices are also widely used but are less useful for comparison with climate model results.

For this analysis we refer to the first EOF of the NH SLP field as the NAO, though it is calculated in an identical fashion to the AO or NAM. EOF is performed using the covariance matrix of the grid point time series for all grid points at latitude $\geq 20^\circ\text{N}$, after they have been adjusted by 1) removing the mean from each time series and 2) multiplying each time series by the square root of the cosine of the latitude of the grid point, as suggested by Chung and Nigam (1999). Contours of the loading factors shown on the figures represent the magnitude of the correlation between the

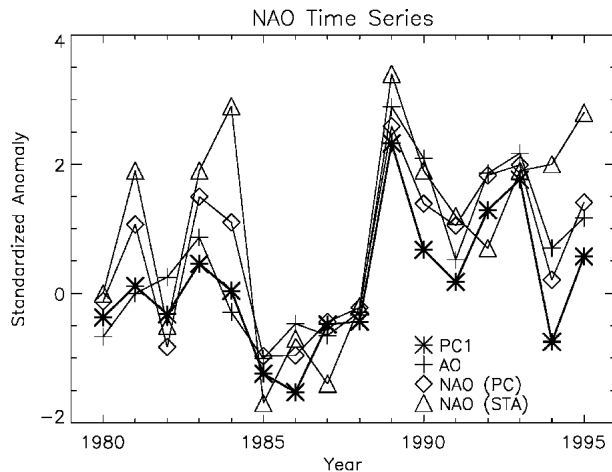


FIG. 1. NAO time series during the AMIP-2 period calculated using several different methods. See Table 1 for explanation of, and cross correlations between, these time series.

EOF and the time series at each grid point. Note that the total variance in the observed field is $\sim 3.1 \times 10^5$ hPa, while the total variance in the model fields ranges from approximately 0.5×10^5 to 6.1×10^5 hPa. While these values are not directly comparable to each other due to the different number of grid points (i.e., different spatial resolutions) included in the analysis, it does indicate that the total variance in the modeled Northern Hemisphere SLP fields covers a range of values, with the observed value somewhere in the middle.

The NAO index calculated for this analysis fluctuates similarly to three other standard NAO indices (Fig. 1) provided in Hurrell and Kushnir (2003). Our index (labeled *PC1* in the figure) is calculated identically to the *AO* index (Thompson and Wallace 1998), except for the time domain over which the EOF was performed. The index labeled *NAO (PC)* is also EOF-based but uses only grid points in the North Atlantic Sector. The index labeled *NAO (STA)* is based on standardized sea level pressure values measured at two stations (see Table 1 for detailed explanation). During the AMIP-2 time period, NAO values were lowest during the mid 1980s and highest in 1989. Correlations between our PC-based index and the standard PC-based indices are ~ 0.9 . Correlations between the station-based index and PC-based indices are lower (~ 0.7 – 0.85 ; see Table 1). Given the high correlations between PC1 and other indices, the results should be robust, regardless of the choice of index.

The principal dataset used for estimating large-scale variations in the spatial extent of NH continental snow cover prior to the late 1990s is based primarily on visible-band satellite imagery. This weekly dataset, produced by the National Oceanic and Atmospheric Administration, covers the period from 1967 to the present (Robinson et al. 1993; Robinson and Frei 2000). Discussions of the applicability of this dataset to climate

TABLE 1. Correlation matrix for winter NAO indices (time series shown in Fig. 1). Values shown with parentheses are detrended correlations. PC1, used in this analysis, is based on the first EOF of NH SLP, DJF, 1979–95; AO is based on the first EOF of reconstructed NH SLP, DJF, 1899–2003; NAO (PC) is based on the first EOF of reconstructed SLP (20° – 70° N; 90° W– 40° E), DJF, 1899–2003; NAO (STATION) is a station-based index based on the difference between normalized sea level pressure (SLP) at Ponta Delgada, Azores, and Stykkisholmur/Reykjavik, Iceland, since 1865. PC1 is calculated based on NCAR–NCEP reanalysis observations (Kalnay et al. 1996); all others taken from Hurrell and Kushnir (2003); also available online at <http://www.cgd.ucar.edu/~jhurrell/nao.html>.

	PC1	AO	NAO (PC)	NAO (STA)
PC1	1.0	0.89 (0.89)	0.92 (0.91)	0.68 (0.64)
AO		1.0	0.84 (0.81)	0.68 (0.63)
NAO (PC)			1.0	0.85 (0.82)
NAO (STA)				1.0

model evaluation and the method used to calculate Eurasian snow extent from AMIP-2 models can be found in Frei et al. (2003). AMIP-2 experiments use observed monthly sea surface temperatures from 1979 through 1995 as the lower boundary condition for the atmosphere (Gates 1992; Gates et al. 1999). These models exhibit improvements over AMIP-1 models in their simulations of the annual cycle as well as interannual variability of continental scale snow extent (Frei et al. 2003). At the regional scale, however, biases in AGCM simulations of snow extent over Eurasia remain significant, and these biases might influence the results presented here (see discussion section). Eurasian snow-cover data are available from the Rutgers University Climate Laboratory Snow Data Resources Center (available online at <http://climate.rutgers.edu/snowcover>).

To compare observed to simulated relationships between Eurasian snow extent and the NAO, a series of EOF and correlation analyses is performed on observed and simulated sea level pressure and snow-cover fields from 14 AMIP-2 AGCMs evaluated in this report (see Table 2) that were available at the outset of this analysis. All calculations were performed with monthly means, which were also averaged to produce seasonal means. Output from all AGCMs are from single members, and all computations using AGCM atmospheric output were done in the models' original grids, except for the computation of pattern correlations for which models were regridded on to the identical $2.5^\circ \times 2.5^\circ$ grid of the Reanalysis data.

3. Results

a. Simulation of NAO pattern

The first EOFs of the SLP fields of AMIP-2 AGCMs consistently display dipole structures in the North Atlantic region that are similar to the observed spatial pattern, although over the Pacific sector intermodel dif-

TABLE 2. Fourteen AMIP-2 AGCMs used in this analysis. Observations are listed on last line. Also listed is the percentage of total NH SLP variance explained by the first EOF, the interannual autocorrelation, and the spatial pattern correlation coefficient between simulated and observed EOF1 coefficients. See text for further explanation.

Acronym	Research institute	Explained variance of EOF1	Interannual autocorrelation	Pattern correlation with NH EOF1
1 CCCMA	Canadian Center for Climate Modeling and Analysis, Canada	38%	-0.16	0.84
2 CCSR	Center for Climate System Research, Japan	49%	-0.16	0.86
3 CNRM	Center National de Recherches Meteorologiques, France	45%	0.01	0.86
4 DNM	Department of Numerical Mathematics, Russia	40%	-0.10	0.75
5 ECMWF	European Centre for Medium-Range Weather Forecasts, United Kingdom	32%	-0.22	0.78
6 JMA	Japanese Meteorological Agency, Japan	41%	0.14	0.76
7 MRI	Meteorological Research Institute, Japan	52%	-0.28	0.91
8 NCAR	National Center for Atmospheric Research, United States	39%	-0.05	0.89
9 PNNL	Pacific Northwest National Laboratory, United States	51%	-0.022	0.85
10 SUNYA	University at Albany, State University of New York, United States	32%	-0.37	0.86
11 UGAMP	The U.K. Universities' Global Atmospheric Modeling Program, United Kingdom	25%	-0.11	0.51
12 UIUC	University of Illinois at Urbana-Champaign, United States	53%	-0.10	0.82
13 UKMO	U.K. Met Office, United Kingdom	39%	-0.13	0.85
14 YONU	Yonsei University, Korea	36%	-0.38	0.82
Observation	NCEP-NCAR reanalysis, United States	39%	0.23	1.00

ferences are perhaps more significant. These results corroborate earlier studies (Delworth 1996; Saravanan 1998; Osborn et al. 1999; Robertson 2001; Stephenson and Pavan 2003) indicating that AGCMs are at least partially capturing the mechanisms responsible for the NAO.

Figure 2 shows the area-weighted pattern correlation coefficients between loading factors (i.e., EOF coefficients) of the first EOF of extratropical NH SLP in AMIP-2 AGCMs and observations. Due to the high spatial correlation in the SLP field, the number of degrees of freedom and, therefore, the appropriate levels of statistical significance are not obvious. We employed two different techniques to estimate the true number of degrees of freedom in the SLP field. The first technique uses the eigenvalues (e) of the PCA of observed extratropical NH SLP. In PCA, which is often used as a data reduction technique, a number of guidelines are often employed to identify components that represent signal rather than noise. Sometimes components with eigenvalues less than 1 are dropped, as this is the average eigenvalue. Other times, a change in slope of the scree plot, which is a plot of the eigenvalues in descending order, indicates a value above which components are retained for further analysis. Although more sophisticated techniques are also used (Wilks 1995; von Storch and Zwiers 1999), the technique employed in our own analysis provides a reasonable estimate of the number of independent signals in this dataset, especially if used as corroboration of a Monte Carlo technique. Only the first five principal components have $e \geq 1$ and a sharp drop in the slope of the scree plot occurs at PC8.

The second technique for estimating statistical significance uses Monte Carlo simulations of red-noise time series with the same lag-1 autocorrelation as the

AO time series from the model simulations. Each artificial time series created from the Monte Carlo simulations is correlated with all the grid points from the AGCM simulated SLP field. Finally the pattern correlations are computed between the observed AO pattern of variability and the generated correlation maps, from which a probability distribution function is derived. (The same technique was first used in Cohen and Saito 2003.) Based on the Monte Carlo simulations, a correlation of 0.76 (0.83) was found to be 95% (99%) significant. This corresponds to about 5 degrees of freedom according to the Student's t test distribution. Therefore based on both techniques we conservatively

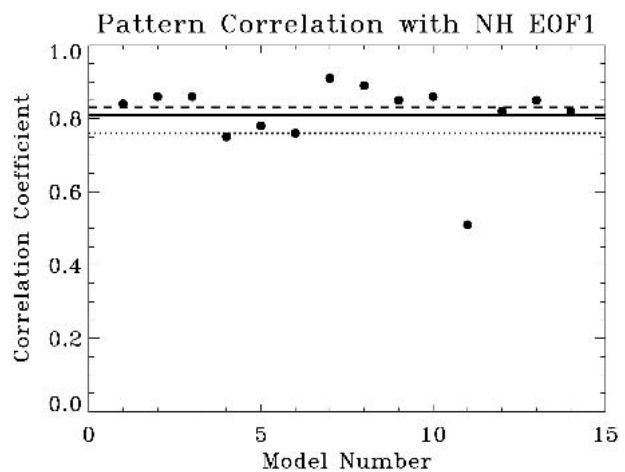


FIG. 2. Pattern correlation coefficient between observed and simulated dominant EOFs of NH SLP. Solid line indicates the model mean, dotted (dashed) line indicates 95% (99%) significance level (see text for explanation of method for estimating statistical significance). Model numbers as shown in Table 1.

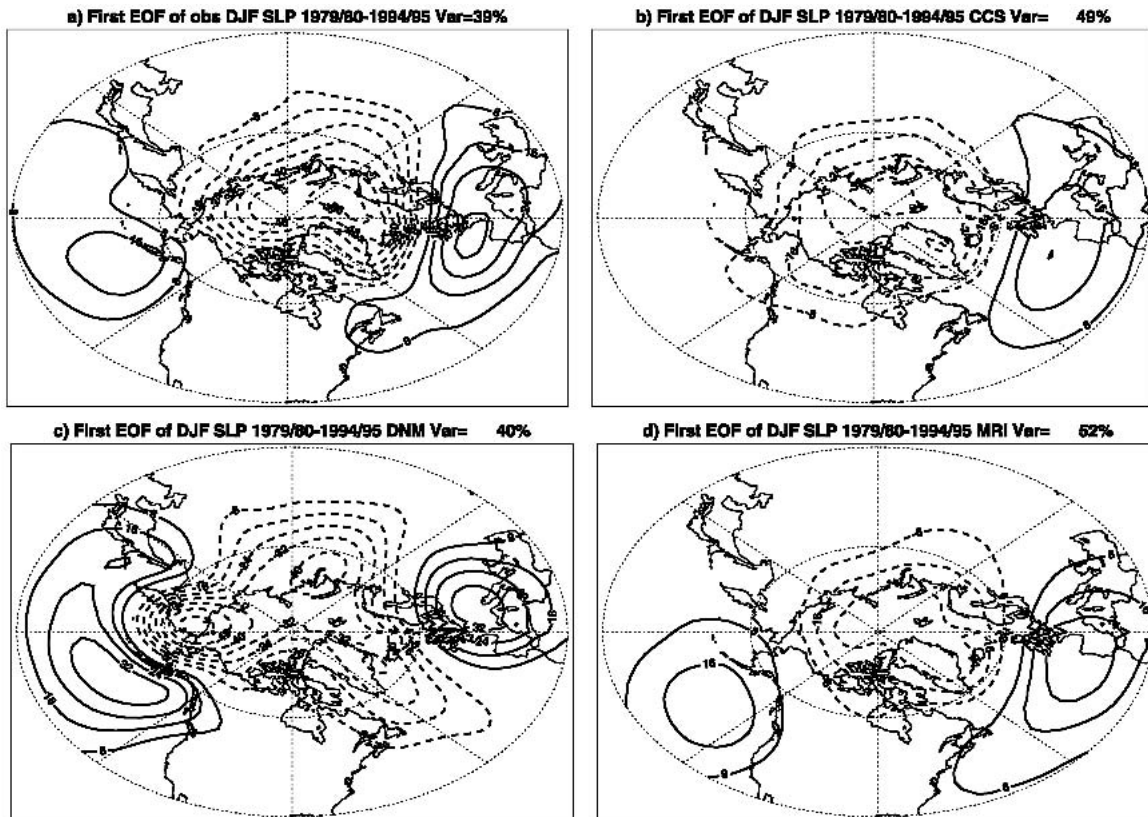


FIG. 3. (a) First EOF of NH SLP (north of 20°N) for Dec, Jan, and Feb for the winters 1979/80–1994/95. (b)–(d) Same as (a) except from simulations by various AGCMs participating in AMIP-2.

estimate that there are five degrees of freedom in the SLP field.

Most pattern correlations are significant, although in one model neither of the two leading EOF patterns resembles the NAO pattern. Other important differences among the models exist. To highlight some of these, in Fig. 3 we present contour plots of the first EOF from observations and from three AGCMs that are representative of the most common differences between models and observations. In the observed pattern (Fig. 3a), an anomaly of one sign covers almost the entire area north of 60°N while an opposite signed anomaly covers the midlatitude ocean basins. Symmetry exists between the North Atlantic and the North Pacific centers of action (at least for the time period covered during the AMIP-2 study). However, the variability is stronger over the North Atlantic. Models often differ from observations by placing too much variability over either the North Atlantic sector [e.g., Center for Climate System Research (CCSR); Fig. 3b] or the North Pacific sector [e.g., Department of Numerical Mathematics (DNM); Fig. 3c]. The latter problem of placing too much variability in the North Pacific sector is probably more egregious, given that during some time periods the coefficients in the region of the North Pacific center of action actually become insignificant

(not shown for observations but very similar to Fig. 3b). The Meteorological Research Institute (MRI) model (Fig. 3d), which has the highest pattern correlation coefficient (see Fig. 2), captures the dipole structure over both ocean basins while appropriately emphasizing the variability over the North Atlantic. The percentage of total variance explained by the first EOF in the models varies between 25% and 53%, with a median value similar to the observed value of 39% (see Table 2). We also compute the interannual autocorrelation coefficients for observed and simulated NAO indices (Table 2). Whereas the observed NAO exhibits a positive autocorrelation, which is indicative of a tendency toward interannual persistence in the phase of the NAO, the models mostly exhibit a negative autocorrelation, which is indicative of a tendency to reverse phase from year to year.

b. Influence of SSTs on NAO variability

Because AMIP-2 models are forced with observed monthly SSTs at the lower boundary, it is reasonable to compare modeled NAO indices to the observed time series. Figure 4 shows temporal correlation coefficients between the observed and simulated NAO time series. Correlation values are generally between 0.0 and 0.3, with a mean of 0.11, although outliers range between

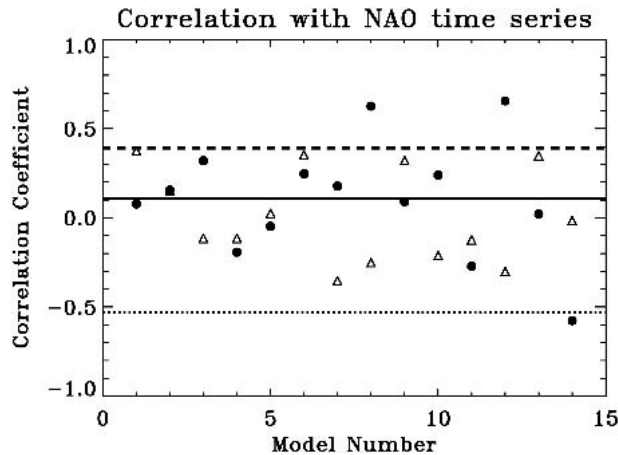


FIG. 4. Correlation between the observed NAO index and the simulated NAO index from 14 AMIP-2 AGCMs (dots); solid line depicts the mean correlation of all 14 AGCMs and dashed line shows correlation of ensemble mean of all 14 model-simulated NAO indices. Also shown are observed (dotted line) and modeled (triangles) lagged correlation coefficients between preceding Oct Eurasian snow extent and the following winter (DJF) NAO. Significance level for $p = 0.05$, two-tailed, $n = 16$ is 0.50.

0.6 and -0.6 . (Detrended time series produce similar results.) These modest values indicate little or no skill in simulating observed interannual variations in the NAO, as no more statistically significant correlations are found than would be expected from a suite of random time series. Doblas-Reyes et al. (2003) found no skill among four AGCMs forced with observed SSTs in predicting the observed NAO index. However, when the four models were merged into a multimodel hindcast, some modest skill was achieved. We did test whether creating a super index by combining the PC time series from all 14 models and correlating it with the observed AO time series improved the results. The correlation value of the mean index is 0.39 (see Fig. 4), which is higher than the mean correlation value of all the models, though it is still not statistically significant. Furthermore, the results do not seem to be robust with respect to the removal of one model from the mean calculation. Therefore, the most reasonable conclusion is that for AMIP-2 models, the influence of SSTs on the phase of the NAO at interannual time scales is insignificant compared to stochastic variability.

The years covered in the AMIP-2 simulations happen to closely coincide with the years used by Hurrell (1995) to demonstrate a robust positive trend in the NAO index (December–March 1980/81–1992/93). Our PC1 NAO index exhibits a positive trend ($+0.12 \text{ yr}^{-1}$), though the significance level is $<90\%$ as measured using the spearman rank correlation method. No interannual trend is apparent in the NAO index constructed from the multimodel hindcast. This suggests that either SSTs are not forcing the positive trend in the NAO observed during the AMIP-2 period or that AGCMs

are as of yet incapable of simulating the observed trend in the NAO when forced solely by SSTs.

c. Eurasian snow cover and NAO variability

The AGCMs examined here all simulate surface snow-cover fields. Interannual variations in the spatial extent of snow at regional to continental scales in the model experiments are uncorrelated to observed values (Frei et al. 2003), as are the simulated NAO indices. Nevertheless, to examine whether any relationship exists between snow extent and the NAO in the simulations, we evaluate correlations between Eurasian snow extent time series and NAO time series that are generated internally to AGCMs, regardless of the lack of temporal correlation to observed values. Lagged correlation coefficients between winter NAO and antecedent Eurasian snow extent from the preceding October in observations and in AMIP-2 AGCMs are shown in Fig. 4. Snow-cover anomalies during October appear to be the most highly influential on winter NAO variability (Cohen et al. 2001), and the observed correlation value between October snow-cover anomalies and the NAO for the AMIP period is -0.53 . In comparison, model correlations are approximately 0.0 ± 0.4 , indicating no consistent relationship among the models and not one model that exhibits behavior similar to observations.

Significant correlations also exist between antecedent Eurasian snow during other months and the subsequent winter NAO. For example, Saito and Cohen (2003) observe significant relationships between summer Eurasian snow cover and the subsequent winter NAO. Figure 5 shows observed and modeled correlation coefficients for each month between leading monthly snow cover and the following winter NAO. The number of significant correlations in the model results is no more than would be expected from the same number of random time series. Observations, however, reveal a pattern of mostly negative correlations from March through July and October through November, with more significant negative correlations during October and November. Detrended correlation analysis of observed time series shows similar results.

4. Discussion and conclusions

The spatial pattern associated with the winter (DJF) NAO is reproduced reasonably well in 13 of the 14 AMIP-2 AGCMs evaluated in this analysis. The ability to simulate the NAO pattern of variability is apparently stable in AGCMs even when boundary conditions are held fixed (Gong et al. 2002) and is consistent with evaluations of NAO simulations in coupled atmosphere–ocean GCMs (Stephenson and Pavan 2003). These results suggest that the mechanisms responsible for the NAO mode of variability in these models are

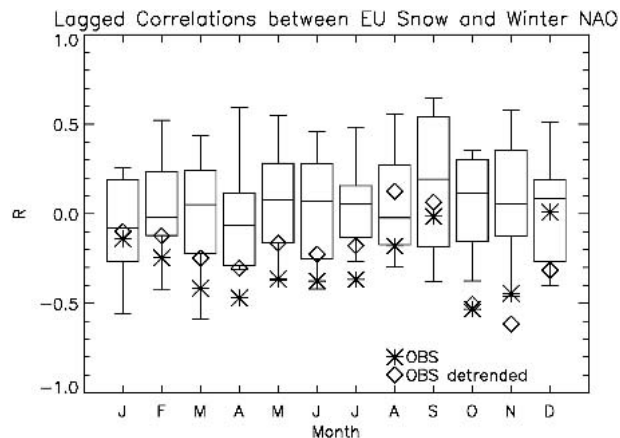


FIG. 5. Lagged correlation coefficients between monthly Eurasian snow cover and the winter NAO (with snow cover preceding the winter NAO). Simulated values by 14 AGCMs shown with box and whisker plots; observed values shown with asterisks. Box plots can be interpreted as follows: middle line shows median value; lower and upper bounds of the box indicate first and third quartiles, respectively; lower and upper whiskers indicate minimum and maximum values, respectively. Eurasian snow cover calculated as in Robinson et al. (1993), available from the Rutgers University Climate Laboratory Snow Data Resources Center (<http://climate.rutgers.edu/snowcover>).

internal to the atmosphere and exist independent of external forcing.

Interannual correlations between NAO indices simulated by AMIP-2 models and observed NAO indices are statistically insignificant, and inconsistent from model to model, despite the fact that models are forced by observed SSTs. This indicates that in AMIP-2 AGCMs the influence of SSTs on the phase of the NAO at interannual time scales is trivial compared to the influence of stochastic variations. This result is consistent with Robertson (2001) and Kushnir et al. (2002) who found that ocean-atmosphere interactions exert little influence on the NAO time series.

It seems likely that AMIP-type model experiments are unsuitable for simulating the processes associated with oceanic boundary forcing of the NAO mode of circulation. The contrast between the observed NAO autocorrelation coefficient (>0) and the modeled autocorrelation coefficients (<0) suggests that there are one or more factors influencing the NAO that have year-to-year memory, and that these factors are neither inherent in the observed SST boundary conditions nor captured by model atmospheric dynamics. Czaja et al. (2003) find that any forcing of NAO variability by the ocean is likely to be on timescales longer than can be adequately addressed during the AMIP-2 time period. Depending on the mechanism(s) responsible for oceanic feedback to the NAO, it may be that coupled GCM experiments are needed to identify those mechanisms and that AMIP-type experiments are inappropriate in this context.

It also appears that no robust relationship between

Eurasian snow extent and the NAO is found in AMIP-2 AGCMs. AMIP-2 AGCMs do not exhibit the same lagged relationship between antecedent Eurasian snow extent and subsequent winter NAO that is observed. At this time we can only speculate on the reasons for the incongruity between the observations and the models, which might be associated with inadequate simulations of either dynamical processes in the troposphere and/or lower stratosphere, or thermodynamic processes at the land surface/atmosphere boundary. For example, in an experiment using the seasonal forecast model at NCEP, Kumar and Yang (2003) found that the influence of snow extent and depth was confined primarily to the lower troposphere, in contradiction to the empirical findings of Saito et al. (2001).

It is also possible that the lack of correlation is due to biases in simulated Eurasian snow extent and depth. While Ye and Bao (2001) and Ye (2001) have found NAO impacts on Eurasian snow depth, to our knowledge no empirical studies have evaluated the impact of snow depth or snow mass on the NAO. However, Gong et al. (2004) found in their AGCM experiments that the snow forcing on large-scale atmospheric variability was sensitive to both snow extent and snow depth. At the continental scale, simulations of Eurasian snow extent by AMIP-2 AGCMs are comparable to observed values in the climatological mean but tend to underestimate interannual variability (Frei et al. 2003). In addition, AMIP-2 AGCMs exhibit significant regional biases over Eurasia. Over eastern China models tend to maintain a snowpack far to the south of the typical southern boundary of the observed snowpack, resulting in an overestimation of snow extent on the order of 10^6 km². Over central Asia, on the other hand, models tend to underestimate snow extent. These biases are consistent across the AMIP-2 AGCMs (Frei et al. 2003). Whether these biases are related to the inability of AMIP-2 AGCMs to capture observed NAO/snow relationships is unknown at this time.

These results are relevant to the debate over the relationship between trends in the NAO and more general climatic changes. Over roughly the past 30 years, including the AMIP-2 time domain, a positive trend greater than would be expected from internal atmospheric variability alone (Feldstein 2002) has been reported in the NAO (Hurrell 1995) and the NAM (Ostermeier and Wallace 2003). What do AGCM experiments reveal about possible future trends of the NAO associated with anthropogenic trace gas emissions? Some studies predict that the upward trend in the NAO will continue with increased greenhouse gas concentrations (Shindell et al. 1999), while others have been less conclusive (Fyfe et al. 1999; Gillett et al. 2000; Zorita and Gonzalez-Rouco 2000). To discriminate between regionalized climate variations associated with circulation patterns such as the NAO and more general variations, the nature of the linkages between boundary conditions and circulation patterns must be clarified, and

the ability of climate models to simulate them must be firmly established. Based on the results regarding two forms of boundary forcing reported here, this latter goal appears to remain elusive.

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