A Large-Ensemble Model Study of the Wintertime AO–NAO and the Role of Interannual Snow Perturbations

GAVIN GONG AND DARA ENTEKHABI

Department of Civil and Environmental Engineering, Massachusetts Institute of Technology, Cambridge, Massachusetts

JUDAH COHEN

Atmospheric and Environmental Research, Inc., Lexington, Massachusetts

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ABSTRACT

Numerous studies have hypothesized that surface boundary conditions or other external mechanisms drive the hemispheric mode of atmospheric variability known as the Arctic Oscillation (AO), or its regional counterpart, the North Atlantic Oscillation (NAO). However, no single external factor has emerged as the dominant forcing mechanism, which has led, in part, to the characterization of the AO–NAO as a fundamental internal mode of the atmospheric system. Nevertheless, surface forcings may play a considerable role in modulating, if not driving, the AO–NAO mode. In this study, a pair of large-ensemble atmospheric GCM experiments (with SST climatology), one with prescribed climatological snow mass and another with freely varying snow mass, is conducted to investigate the degree to which the AO–NAO is modulated by interannual variability of surface snow conditions. Statistical analysis of the results indicates that snow anomalies are not required to produce the AO–NAO mode of variability. Nevertheless, interannual variations in snow mass are found to exert a modulating influence on the AO–NAO mode. Snow variations excite the AO pattern over the North Atlantic sector, produce correlated hemispheric AO features throughout the troposphere and stratosphere, and generate autumn sea level pressure anomalies over Siberia that evolve into the winter AO–NAO. These numerical modeling results are consistent with previous observational analyses that statistically link the AO–NAO mode with the Siberian high and associated snow cover variations.

1. Introduction

The dominant pattern of Northern Hemisphere climate variability is characterized by simultaneous and opposite-signed oscillations of atmospheric mass between high and midlatitudes. The oscillations are most prevalent in the winter season, and occur over a wide range of timescales, from intraseasonal to interdecadal. This pattern of climate variability is commonly referred to hemispherically as the Arctic Oscillation (AO; Thompson and Wallace 1998) or the Northern Annular Mode (NAM; Thompson and Wallace 2001), and regionally as the North Atlantic Oscillation (NAO; Wallace and Gutzler 1981). The NAO in particular has been shown to exert a strong influence on climate in western Europe and eastern North America, via latitudinal shifts in the wintertime North Atlantic storm track and associated variations in temperature, precipitation, and cyclonic activity (Hurrell 1995; Serreze et al. 1997). Climatic conditions have a considerable societal impact in the populous regions bordering either side of the North Atlantic. This has prompted research efforts to better understand the mechanisms that drive this overall pattern of variability (hereafter called the AO–NAO). A major goal of these efforts is to predict the phase and magnitude of the AO–NAO pattern on seasonal timescales, in order to better anticipate wintertime climatic conditions over the midlatitude Northern Hemisphere.

Corresponding author address: Dara Entekhabi, Department of Civil and Environmental Engineering, MIT Bldg. 48-331, Cambridge, MA 02139.
E-mail: darae@mit.edu.

2. Background

The ocean boundary has received considerable attention as a potential driving force for the observed AO–NAO pattern in midlatitude, Northern Hemisphere winter climate. A number of studies have detected significant relationships between observed interannual SST variations and the NAO (e.g., Latif et al. 2000; Rodwell et al. 1999; Robertson et al. 2000; Dong et al. 2000; Hoerling et al. 2001). Different studies have linked climate in the North Atlantic sector atmosphere with SSTs in the North Atlantic, South Atlantic, and equatorial Pacific. Seager et al. (2000) found that North Atlantic SSTs do not force, but rather respond to, changes in atmospheric circulation. Unlike the El Niño–Southern Oscillation (ENSO), which has been linked with climate variability in the Tropics and even the extratropics, a definitive relationship has yet to be established between SSTs and climate variability in the North Atlantic sector.

Another set of studies links the surface and upper-level features of the AO–NAO mode of variability to planetary-scale tropospheric–stratospheric wave interactions (e.g., Baldwin and Dunkerton 1999; Christiansen 2000; Kodera and Kuroda 2000; Perlwitz et al. 2000). These studies portray the equivalent barotropic nature of the AO–NAO mode, extending from the surface to the midstratosphere. However, different studies have found signals to originate in both the troposphere and stratosphere, and both upward and downward wave propagations have been detected. Furthermore, no clear causal mechanisms have been linked to the signal origins. Anthropogenic climate change factors and episodic events such as volcanic eruptions have been suggested (Kodera and Yamazaki 1994; Graf et al. 1994; Shindell et al. 1999), but have yet to prove conclusive.

Land surface characteristics also have the potential to affect Northern Hemisphere climate variability, since the largest landmasses on earth reside in the midlatitudes of the Northern Hemisphere. Furthermore, much of the wintertime land surface is subject to considerable variations in seasonal snow cover and snow mass. Over vast contiguous land surface areas such as Eurasia and North America, the change in diabatic heating caused by snow anomalies may affect not only local surface temperature (Leathers and Robinson 1993), but also atmospheric dynamics and thermodynamics, and consequently regional and remote climatic conditions (Cohen 1994; Walland and Simmonds 1996). The recognized relationship between Eurasian winter snow cover and subsequent Indian summer monsoon rainfall (Hahn and Shukla 1976; Barnett et al. 1989; Douville and Royer 1996; Bamzai and Shukla 1999) demonstrates the ability of Eurasian snowfall anomalies to have a significant impact on large-scale climate variability.

Despite this potential, possible linkages between interannual snow anomalies and the AO–NAO pattern of variability have received relatively little attention. Walland and Simmonds (1997) and Cohen and Entekhabi (2001) demonstrate the impact of winter hemispheric snow anomalies on climatic conditions in the North Pacific and North Atlantic basins, respectively. Watanabe and Nitta (1998, 1999) attribute the decadal-scale shift in winter Northern Hemisphere climate that occurred in 1988/89 to a large negative anomaly in Eurasian snow cover that occurred in the preceding autumn. Cohen and Entekhabi (1999), Cohen et al. (2001), and Saito et al. (2001) make the case that the leading mode of winter Northern Hemisphere climate variability (i.e., the AO–NAO) is significantly correlated to the development of the surface Siberian high pressure system during autumn, and associated snow anomalies. They argue that the observed seasonally lagged correlation is indicative of autumn snow anomalies driving winter climate, and hypothesize several teleconnection pathways linking Eurasian snow to the climate in the North Atlantic sector.

The aforementioned studies utilize either statistical analyses of observed long-term datasets, or numerical simulations of atmospheric general circulation models (GCMs). Both types of analyses pose challenges to identifying surface boundary conditions or other mechanisms which drive Northern Hemisphere climate variability. One difficulty with observational analyses is the interdependency of numerous parameters in the complex atmospheric system, which make it problematic to associate causality with any observed statistically significant relationship. GCM studies provide an experimental platform for isolating the climatic effect of a specified forcing. However, despite continual advances, GCMs still do not replicate the atmospheric system with adequate precision and accuracy, and simulation results can vary substantially between different models. In addition, GCM studies must either be of sufficiently long duration, or include a sufficient number of independent realizations, to distinguish the climatic response to a boundary forcing from intrinsic model climate variability. In the case of the GCM studies involving snow forcing described above, the exploratory nature of these investigations has generally limited them to short-term integrations (1–6 months), and five or fewer realizations.

Previous studies have attempted, with varying degrees of success, to establish that interannual variations in a hypothesized surface boundary condition or other physical mechanism serves as the principal driving force behind the AO–NAO mode of variability. Although SSTs, land surface snow, and upper-level atmospheric conditions have all demonstrated significant statistical relationships with the AO–NAO, no single feature has emerged as the dominant characteristic that precedes, governs, and enables the prediction of winter climate in the North Atlantic sector. Hence, recent studies have suggested that the AO–NAO is not an externally forced mode at all, but rather a fundamental internal mode of the atmospheric system (Robertson 2001; Baldwin 2001). Feldstein (2000, 2002) uses daily unfiltered data to demonstrate that the AO–NAO fluctuates primarily on intraseasonal timescales arising from processes internal to the atmosphere, and that on interseasonal timescales variations in the AO–NAO may be considered simply ''climate noise.''

A related question that therefore arises is, if the AO–NAO is in fact not driven by external forcings, then can it still be modulated by boundary conditions on an interseasonal timescale at a level greater than internal noise? This study addresses this issue, with a focus on interannual snow cover variations as a modulating boundary condition. A pair of large-ensemble atmospheric GCM experiments, one with prescribed climatological snow mass and another with freely varying snow mass, is conducted to investigate the degree to which wintertime extratropical Northern Hemisphere climate variability is driven by interannual variability in surface snow conditions. Section 3 describes the GCM that is used, the experimental design, and the external boundary forcings that are applied in the two simulations. Section 4 assesses the internal or externally forced nature of the two model-simulated AO–NAO modes of variability. Section 5 evaluates whether interannual snowmass variations can at least modulate, if not force, the AO–NAO. Conclusions are presented in Section 6.

3. Experimental design

Numerical simulations are conducted using the ECHAM3 GCM, developed by the Max Planck Institute for Meteorology in Hamburg, Germany. ECHAM3 has evolved from the spectral operational weather forecast model used at the European Centre for Medium-Range Weather Forecasts (ECMWF), incorporating physical parameterizations and revised numerical methods appropriate for climate simulations (Roeckner et al. 1992). In general, ECHAM3 simulates observed climate features with considerable skill, and the model has been used in a number of climate-related studies (Kaurola 1997). ECHAM3's representation of snow cover and snow mass compares reasonably well with observations and with other GCMs (Foster et al. 1996). All model integrations are performed with the spherical harmonic series triangularly truncated at wavenumber 42 (T42), which corresponds to a Cartesian resolution of roughly 2.8° latitude by 2.8° longitude. Vertical discretization consists of a second-order finite difference scheme applied to a 19-layer hybrid sigma-pressure coordinate system, with the uppermost layer centered at 10 hPa.

Two numerical experiments are conducted, each consisting of 20 independent integrations of autumn/winter (September to February) climate. A set of 20 independent 1 September initial conditions is used to start the 6-month model integrations, obtained from a 20-yr control integration of the base model. Initiating the numerical experiments with 1 September conditions from 20 different years instead of 20 consecutive days within a single year ensures that the 20 model integrations are reasonably independent. The same set of initial conditions is applied to both experiments. Although we are primarily interested in boreal winter climate, the model integrations include boreal autumn months because of the suggested link between autumn snow cover and winter climate (Cohen and Entekhabi 1999).

The first experiment is designated as FIX. It consists of prescribed monthly climatological SSTs and sea ice throughout the global model domain, derived from the Center for Ocean–Land–Atmosphere Studies/Climate Analysis Center (COLA/CAC) dataset developed for the Atmospheric Model Intercomparison Project (Gates 1992). Climatological snowmass values are also prescribed, on a weekly basis throughout the global model domain. The "fixed" snowmass climatology applied in this experiment is derived by averaging the weekly model-generated snow mass from the 20-yr control run. This experiment, therefore, includes no interannual variations in either the land surface snow or the SST boundary condition.

The second experiment is designated as FREE. Monthly climatological SSTs and sea ice are once again prescribed. Snow mass is not prescribed, but left unaltered as an internally varying, that is, "free," land surface state variable. This experiment therefore includes intraseasonal and interannual variations in the land surface snowmass boundary condition as generated by the model, but no interannual variations in the SST boundary condition. To demonstrate the extent of these interannual snow variations, Fig. 1 presents the Northern Hemisphere snowmass field for two realizations of the FREE simulation, which generate relatively extensive (Fig. 1a) and limited (Fig. 1b) snow. The figures represent mid-October conditions, when interannual variations in snow mass are expected to be substantial. The most notable differences occur over southern Siberia and southern Canada.

These numerical experiments deal with the specification of SST and snow cover/snowmass boundary conditions in the model. Whereas SSTs represent true passive boundary conditions for this atmospheric model experiment, snow is treated by the model as an active internal state variable, since snow accumulation, snowmelt, and a five-layer land surface scheme are included in the model's land surface parameterization. Thus specifying snow mass as an external boundary condition in the FIX experiment may potentially disrupt the energetics of the land surface. To minimize any surface imbalances, snow mass is specified at each time step prior to updating the surface moisture and energy fluxes, so that the prescribed snow mass is subject to melting and evaporation. The resulting surface energy balances for these experiments were evaluated and compared, which confirmed that the energy balance terms were not unduly disrupted in the FIX experiment.

4. AO–NAO mode of variability

The leading mode of variability over the 20 realizations of the model-simulated Northern Hemisphere winter [December–January–February (DJF) average] is evaluated using empirical orthogonal function analysis. This measure captures the dominant spatial patterns of temporal variability within a gridded dataset. Figure 2 presents the leading empirical orthogonal function (EOF1) for the Northern Hemisphere (north of the equator) sea level pressure (SLP) field, for the FIX simulation. The characteristic dipole pattern between high and midlatitudes associated with the hemispheric AO mode is produced and explains 43% of the total variance in the SLP field, which is comparable to that obtained

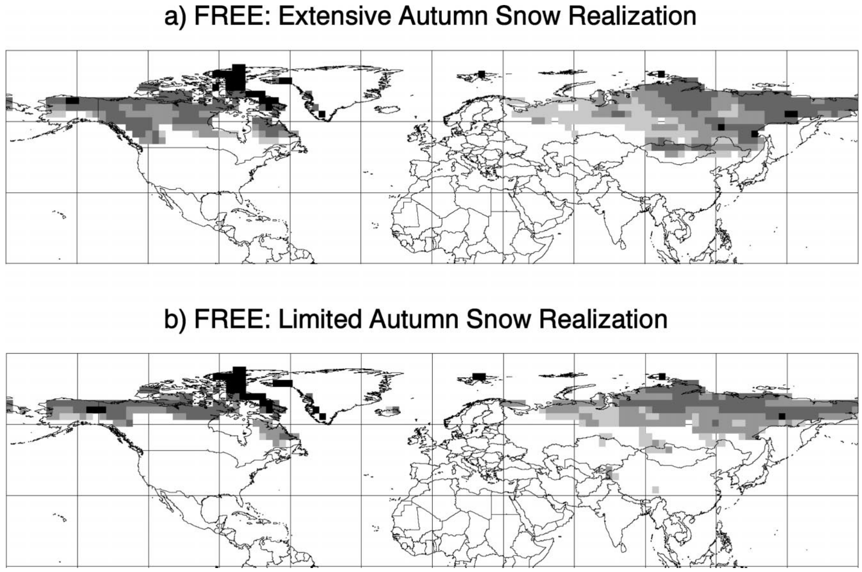

FIG. 1. Mid-Oct Northern Hemisphere snowmass fields for the FREE simulation: (a) realization with extensive autumn snow; (b) realization with limited autumn snow. Shading represents 0.1–1 cm (lightest), 1–5 cm, 5–25 cm, and 25+ cm (darkest).

from the National Centers for Environmental Prediction–National Center for Atmospheric Research (NCEP–NCAR) reanalysis data (Cohen and Entekhabi 2001). However, in Fig. 2 the Pacific anomaly is stronger than the Atlantic anomaly, whereas observations generally indicate a stronger anomaly in the Atlantic. The corresponding EOF1 for the 500-hPa geopotential height field (not shown) is likewise comparable to observations.

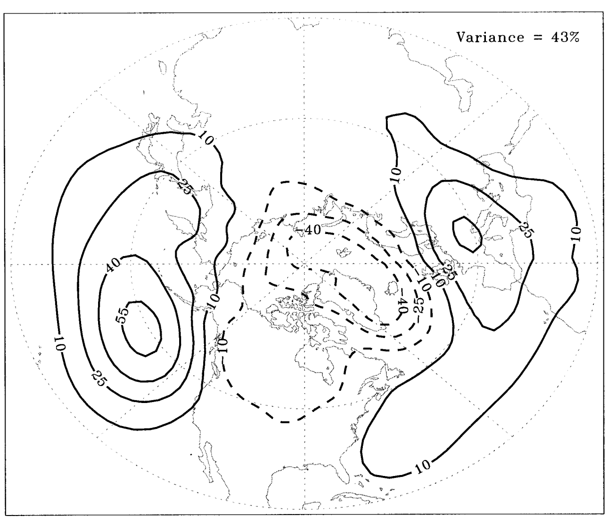

FIG. 2. Leading EOF1 of Northern Hemisphere DJF average SLP field, computed from the 20 realizations of the FIX simulation (prescribed monthly climatological SSTs and prescribed weekly climatological snow mass), including the percent of total variance in SLP explained by the EOF1 mode. Contour values are arbitrary, and represent relative magnitudes within the spatial EOF1 pattern.

Figure 2 indicates that even in the absence of interannual variations for both the snowmass and SST boundary forcings, the model climate still exhibits the classic AO–NAO pattern of variability at the surface. Since interannual external forcings are not included in the FIX simulation, the resulting AO–NAO mode arises solely from intraseasonal fluctuations, that is, internal climate noise (Feldstein 2002). This result agrees with recent studies (described in section 2) that assert that the principal mechanisms that drive this dominant mode of climate variability likely do not reside in the surface boundary conditions, but rather are internal to the atmospheric system. This internal mode is also observed in other GCMs forced with seasonally varying climatological SSTs (Robertson 2001), and in coupled atmospheric–ocean models (Fyfe et al. 1999; Stone et al. 2001).

Although surface conditions do not appear to govern the AO–NAO, they may nevertheless be influential enough to modulate this pattern of climate variability in some respect. Both SSTs and snow mass have been shown to influence other features of the climate system, such as ENSO-related climate variability and summer monsoon activity. Also, previous studies have revealed statistically significant relationships between the AO–NAO and both SSTs and snow mass. Even if the AO–NAO is a fundamental internal atmospheric mode, it can potentially be varied by surface boundary conditions, such that the total interannual variability in the observed AO–NAO mode is not composed entirely of internal climate noise. In the next section, one such forcing, interannual snowmass variations, will be studied in detail.

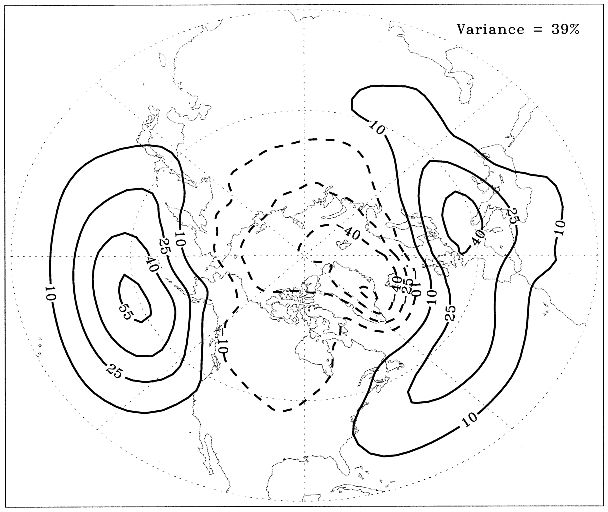

FIG. 3. Leading EOF1 of Northern Hemisphere DJF average SLP field, computed from the 20 realizations of the FREE simulation (prescribed monthly climatological SSTs and internally varying model snow mass), including the percent of total variance in SLP explained by the EOF1 mode. Contour values are arbitrary, and represent relative magnitudes within the spatial EOF1 pattern.

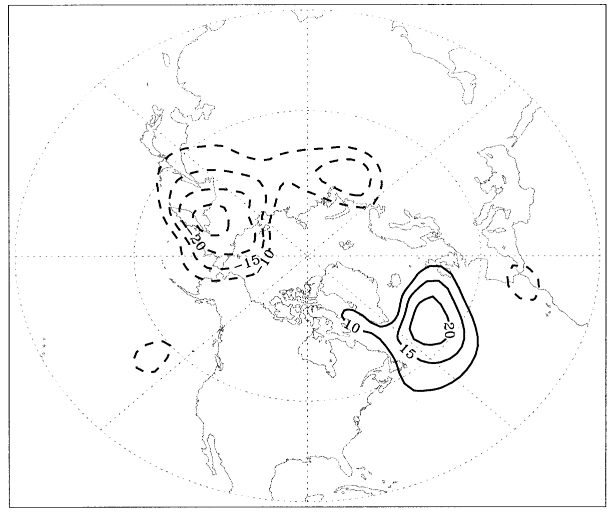

FIG. 4. Difference in leading EOF1 of Northern Hemisphere DJF average SLP field, between the FREE and FIX simulations. Contour values are arbitrary and represent relative magnitudes within the spatial EOF1 difference pattern.

5. Effect of interannual snowmass variations

Before evaluating the effect of interannual snowmass variations on the characteristics of the modeled AO–NAO mode, the impact of prescribing climatological snowmass values on the mean climatic state of the model should be addressed. Ensemble average Northern Hemisphere gridpoint values over all 20 realizations are compared between the FREE and FIX simulations. This comparison is made for a number of land surface and climate state variables, including surface albedo and temperature, SLP, geopotential heights, and zonal wind (not shown). Surface albedo and temperature fields do not exhibit any significant difference, indicating a negligible ensemble mean local thermodynamic response to interannual snow variability. Winter SLP, geopotential height, and zonal wind fields do exhibit some regions of significant difference between the FREE and FIX simulations, particularly at high to midlatitudes. These differences are somewhat reminiscent of the AO–NAO mode, although the spatial patterns are not well organized. This slight shift in mean climatic state might suggest that winter climate responds nonlinearly to the positive versus negative snow anomalies contained in the FREE simulation, relative to the FIX simulation. On the other hand, it may simply be a random consequence of naturally occurring internal atmospheric variability. The mean climatic response to forced positive and negative snow anomalies is the subject of ongoing research (see the conclusions).

a. AO–NAO characteristics

Figure 3 presents the leading EOF1 of the Northern Hemisphere (north of the equator) SLP field, for the FREE simulation. As for the FIX simulation in Fig. 2, the major AO characteristics (i.e., spatial pattern and percentage of total variance explained) are produced, although the Pacific anomaly is once again stronger than the Atlantic anomaly. In addition, the expansion coefficients for the two EOF1 patterns have statistically identical standard deviations (87.6 hPa for FIX, 85.7 hPa for FREE), indicating that the interannual snow variations in the FREE simulation do not result in increased hemispheric climate variability. These similarities are further demonstration that the AO–NAO is a fundamental internal mode of the atmosphere. However, a close comparison of the FIX and FREE simulations indicates that the inclusion of interannual snow variations results in subtle but notable changes to the AO pattern. These changes are revealed in Fig. 4, which presents the difference between the EOF1 fields shown in Figs. 2 and 3. Interannual snow variations result in relatively stronger climate variability over Siberia, a broad land surface region subject to considerable snow and snow variations. Climate variability is also relatively stronger over Greenland, Iceland, the North Atlantic, and western Europe, regions far removed from major interannual snow variations, but typically associated with the NAO. It should be noted that the contour intervals in Fig. 4 are less than in Figs. 2 and 3. Thus interannual snow variations enhance the AO mode of variability (albeit slightly) only over specific regions that complement recent observational studies associating snow cover Eurasia with North Atlantic climate (Cohen and Entekhabi 1999; Cohen et al. 2001; Saito et al. 2001).

The expansion coefficients for the EOF1 patterns shown in Figs. 2 and 3 also exhibit a very low and statistically insignificant correlation (0.18) between the

FIX and FREE simulations. Similarly, a regional NAO index is computed, by differencing winter SLP values between the Arctic and North Atlantic centers of action indicated by the EOF1 modes. As for the EOF1 expansion coefficients, this NAO index over all 20 realizations is poorly correlated (0.12) between the FIX and FREE simulations. This lack of correlation may be associated with the introduction of interannual snow variations in the FREE simulations, under the presumption that a significant correlation would result if the snow anomalies had a negligible effect on climate variability. On the other hand, the poor climate index correlation between the realizations of the two experiments may simply be attributed to chaos in the system.

In an attempt to confirm this reasoning, the Antarctic Oscillation (AAO) index is evaluated, which describes a pattern of high- to midlatitude Southern Hemisphere climate variability analogous to the NAO index in the Northern Hemisphere (Gong and Wang 1999). The winter AAO index is much more highly correlated (0.52, which is statistically significant at 95%) between the FIX and FREE simulations than is the winter NAO index (0.12). Due to the very limited occurrence of land surface snow in the modeled Southern Hemisphere, interannual snow variations are not expected to have a substantial impact on Southern Hemisphere climate variability. Therefore the significant correlation in the southern index suggests that the lack of correlation in the northern index may be due in part to the considerable snow variations in the Northern Hemisphere and associated climate modulations. It must be noted that this exploratory evaluation of the AAO index is not a precise test of the predictability of the AO–NAO and AAO, nor of the implications of the poor correlations found for the AO–NAO indices. Nevertheless, this comparison between Northern and Southern Hemisphere fields yields results that are notable and that may provide insight into the role of the differences between the experiments, that is, the results maintain the possibility that interannual snow variations are a modulator of winter Northern Hemisphere climate.

b. Vertical extent of the AO–NAO

An important facet of the AO as derived from observed data is its vertical extent, characterized by highly correlated dominant modes of variability extending from the surface into the stratosphere (Thompson and Wallace 1998). This feature is evaluated for both the FIX and FREE simulations by correlating the winter (DJF) EOF1 expansion coefficients between SLP and geopotential height at various atmospheric pressure levels, as shown in Fig. 5. For the FREE simulation, the surface signal is well correlated with the troposphere, and reasonably well correlated throughout most of the stratosphere. Statistically significant correlations at 95% extend to a height of 20 hPa, which is consistent with observations. This result implies that with the inclusion of interannual snow variations, the resulting AO–NAO mode of variability occurs throughout the atmosphere. However, for the FIX simulation, even though the correlation with the troposphere is again very strong, the correlation with the stratosphere is notably weaker, and no longer statistically significant. This result is strongly suggestive that without interannual snow variations, the dominant modes of variability in the troposphere and stratosphere may be essentially uncoupled. Thus, interannual snow variations at the land surface boundary are necessary to maintain the full vertical extent of the AO–NAO mode of variability. Hypothesized dynamical mechanisms for generating this connectivity include the vertical propagation of Rossby waves, excited by surface diabatic heating changes over snow anomalies (Saito et al. 2001; Cohen et al. 2002).

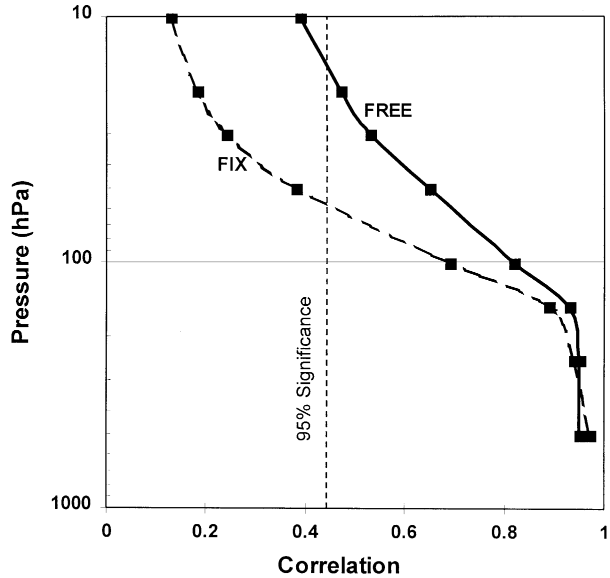

FIG. 5. Northern Hemisphere DJF average EOF1 expansion coefficient correlation between SLP and geopotential height at various pressure levels throughout the troposphere and stratosphere, for the FREE (solid line) and FIX (dashed line) simulations.

c. Origins of the AO–NAO

In Cohen et al. (2001), 27 yr of NCEP–NCAR reanalysis data (Kalnay et al. 1996) are used to correlate a winter climate index describing the AO–NAO mode of variability to 45-day-average gridded values of surface parameters (e.g., SLP). A temporal sequence of correlation fields from October to January demonstrates the role of the Siberian high in Northern Hemisphere climate variability. In this section, a similar analysis is conducted using GCM output from both the FREE and FIX simulations, intended as a numerical modeling counterpart to the observational analysis of Cohen et al. (2001).

Figure 6 presents the correlation between the EOF1 expansion coefficients for winter (DJF) Northern Hemi-

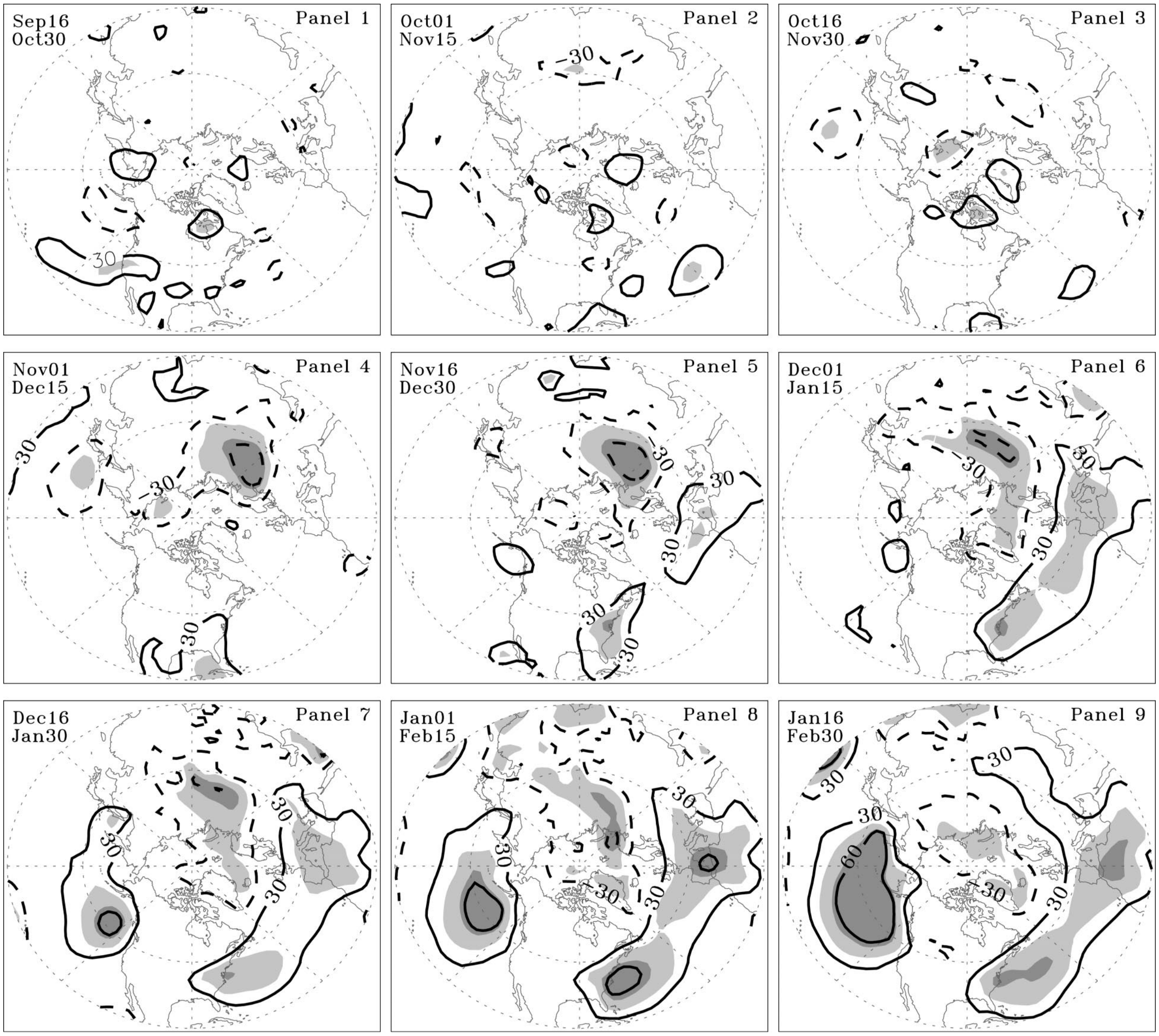

FIG. 6. Percent correlation between EOF1 expansion coefficients for DJF average Northern Hemisphere 50-hPa geopotential height field and a series of 45-day-average gridpoint SLP, for the FREE simulation. Contours drawn at ±30%, 60% and 80% correlation. Solid (dashed) lines denote positive (negative) correlation. Light (dark) shading represents absolute correlations in excess of 44% (56%), representing 95% (99%) statistical significance.

sphere 50-hPa geopotential height, and a series of gridded 45-day-average SLP fields spanning the September to February GCM integration period, for the FREE simulation. For reference, regions of 90% and 95% statistical significance as determined by *t* tests are indicated by light and dark shading, respectively. It should be noted that such traditional measures of statistical significance have come under recent scrutiny (Nicholls 2000); therefore, the focus should be on the overall spatial patterns of correlation, and not just on regions of statistical significance. Figure 6 indicates a region of negative correlation that emerges in western Siberia in late autumn (panel 4), and expands northward with the onset of winter into the Arctic and high-latitude North Atlantic. Concurrent with this expanding region of negative correlation during winter is the emergence of a region of positive correlation in the midlatitude North Atlantic and western Europe (panel 6), and another positive correlation region in the North Pacific (panel 7). The ultimate late winter correlation field (panels 8 and 9) resembles the AO–NAO pattern of variability.

Figure 6 indicates that with the inclusion of interannual snow variations, the winter AO–NAO signal originates in the autumn as a SLP anomaly over Siberia, which subsequently migrates over the Arctic and into the North Atlantic during the course of the autumn/winter season. As a result, the winter Icelandic and Aleutian low pressure cells are forced to migrate southward, contributing to the dipole SLP anomaly pattern characteristic of the AO–NAO. Figure 6 is analogous

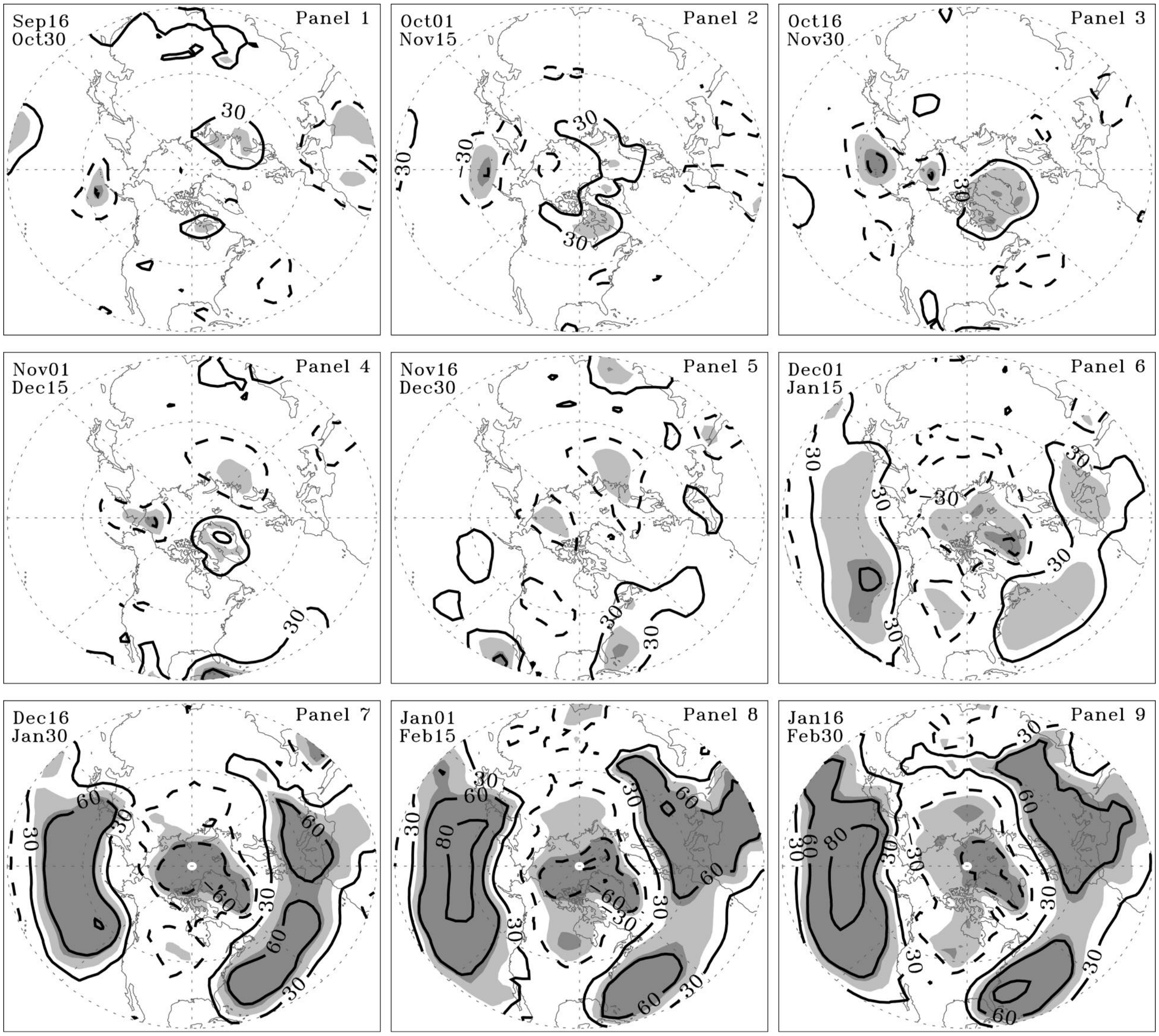

FIG. 7. Percent correlation between EOF1 expansion coefficients for DJF average Northern Hemisphere SLP field and a series of 45-day-average gridpoint SLP, for the FREE simulation. Contours drawn at $\pm 30\%$, 60%, and 80% correlation. Solid (dashed) lines denote positive (negative) correlation. Light (dark) shading represents absolute correlations in excess of 44% (56%), representing 95% (99%) statistical significance.

to Fig. 2 (based on 27 yr of NCEP–NCAR reanalysis data) of Cohen et al. (2001), which similarly portrays a surface teleconnection pathway originating as an autumn SLP anomaly over Siberia, and migrating over the Arctic to contribute to the winter AO–NAO pattern. Note that Siberia is a broad high- to midlatitude land surface region, whose dominant autumn/winter land surface feature is snow cover. In addition, interannual snow cover variability over Siberia is most likely to occur in autumn, as opposed to winter when snow cover is fully established. Thus the origin of the AO–NAO pattern is coincident with the region and season of greatest interannual snow variability in the FREE simulation, which demonstrates how autumn snow conditions in Siberia may modulate winter climate throughout the extratropical Northern Hemisphere. This apparent surface pressure teleconnection pathway implies that even though the AO–NAO mode in the FREE simulation is generated from intraseasonal fluctuations in the atmosphere (i.e., internal climate noise), it is not isolated from interseasonal variations in surface boundary conditions.

Figure 6 also provides further evidence that snow variations modulate climate patterns throughout the atmosphere, since the surface SLP teleconnection pathway for the AO–NAO signal is observed using EOF1 expansion coefficients at 50 hPa, that is, an upper-level climate index. Figure 7 is identical to Fig. 6, except that the gridpoint SLP fields are instead correlated to EOF1 expansion coefficients for winter (DJF) SLP, the traditional surface level AO climate index. Figure 6 indicates an obvious AO–NAO winter correlation pattern (panels 7–9), since the climate index used is derived from SLP values, so that winter gridpoint SLP values are essentially correlated to themselves. However, the autumn Siberian SLP anomaly and subsequent Arctic migration (panels 4–6) are not as prevalent using the surface index in Fig. 7 as with the upper-level index in Fig. 6. Consistent with the statistically significant correlation of EOF1 expansion coefficients throughout the troposphere and stratosphere in the FREE experiment (Fig. 5), this result suggests that a vertical teleconnection pathway may also exist.

Both Figs. 6 and 7 correlate gridpoint SLP values to winter (DJF average) EOF1 climate indices. However, this results in a temporal overlap between the winter index and the 45-day-average periods beginning with panel 4 of the figures. The Siberian SLP anomaly associated above with the AO–NAO signal first emerges during an overlapping period (panel 4), which raises the possibility that the observed Siberian SLP anomaly is an artificial result of the overlapping periods. Therefore, the correlation sequences shown in Figs. 6 and 7 were reevaluated using EOF1 expansion coefficients averaged over January–February only, so that overlaps do not occur until panel 6. This 2-month period was sufficiently long to exhibit 50-hPa geopotential height and SLP EOF1 patterns equivalent to those for the DJF average. The resulting correlations sequences (not shown) are very similar to, and thereby validate, Figs. 6 and 7; a negative SLP anomaly again appears over Siberia in autumn, and migrates over the Arctic to produce the winter AO–NAO pattern.

Figure 8 shows the gridpoint SLP correlation sequence to winter 50-hPa geopotential height EOF1 expansion coefficients, computed for the FIX simulation. In contrast to the FREE simulation (Fig. 6), an autumn SLP anomaly over Siberia fails to materialize without interannual snow variability. Furthermore, the winter AO–NAO pattern also does not materialize, since for the FIX simulation the stratospheric EOF1 mode was found to be uncoupled from the surface AO–NAO signal (see Fig. 5). Comparison of Fig. 8 (FIX simulation, upper-level index) with Fig. 6 (FREE simulation, upper-level index) clearly demonstrates the manner in which snow variations modulate the interannual variability of the AO–NAO. Exclusion of snow variations (FIX) results in an internal AO–NAO mode driven purely by intraseasonal fluctuations (i.e., climate noise), which is limited to the lower atmosphere, and unrelated to autumn conditions over Siberia. Inclusion of interannual snow variations (FREE) results in a modulated AO–NAO mode influenced by varying surface boundary conditions, coupled to the stratosphere, and that originates as an autumn SLP pressure anomaly over Siberia.

6. Conclusions

The objective of this study is to evaluate the extent to which interannual variations in the land surface snow boundary condition can trigger, excite, or otherwise modulate the fundamental internal AO–NAO mode of variability. Large-ensemble (20) GCM experiments conducted with and without interannual snow variability are compared to evaluate changes in Northern Hemisphere climate characteristics. Even though interannual snow variations do not drive the AO–NAO, results indicate that they are influential enough to alter regional and temporal aspects of the overall AO–NAO from the pattern that arises purely from intraseasonal climate noise, in several ways.

- The AO mode of variability over the North Atlantic sector is apparently enhanced by interannual snow variations, as indicated by relatively stronger winter SLP EOF1 anomalies over Greenland, Iceland, the North Atlantic, and western Europe.
- As evidenced by significant correlation of the EOF1 expansion coefficients between atmospheric levels, the AO–NAO mode of variability extends into the stratosphere only when snow variations are included, otherwise the simulated AO–NAO pattern is limited to the troposphere.
- With the inclusion of interannual snow variations, the interannual winter AO–NAO signal is found to originate in autumn over Siberia, a season and region of maximum snow cover variability. Without interannual snow variations, the interannual winter AO–NAO signal may be unrelated to autumn conditions over Siberia.

These modeling results are consistent with previous observational studies in which a causal relationship between Eurasian snow cover and winter extratropical Northern Hemisphere climate variability has been hypothesized and investigated. The required inclusion of interannual snow variability to reproduce the full vertical extent of the AO–NAO mode is suggestive of a vertical teleconnection pathway, as discussed in Saito et al. (2001). The autumn SLP anomaly over Siberia, which evolves into the winter AO–NAO pattern, echoes the surface teleconnection pattern described in Cohen et al. (2001). This agreement between observational analyses and numerical modeling experiments is critical to building a more complete case regarding the ability of interannual snow variations to at least modulate, if not drive, midlatitude Northern Hemisphere climate variability during the winter season. The results of this study and its predecessors consistently suggest that anomalous values of the interannual winter AO–NAO index may be preceded by anomalous autumn snow conditions in Siberia.

Despite the apparent ability of interannual snow variations to modulate the AO–NAO pattern of variability, and the identification of Siberia as a possible source region for this modulation, winter climate indices such as the SLP EOF1 expansion coefficients and the NAO index are not directly correlated with the autumn snow

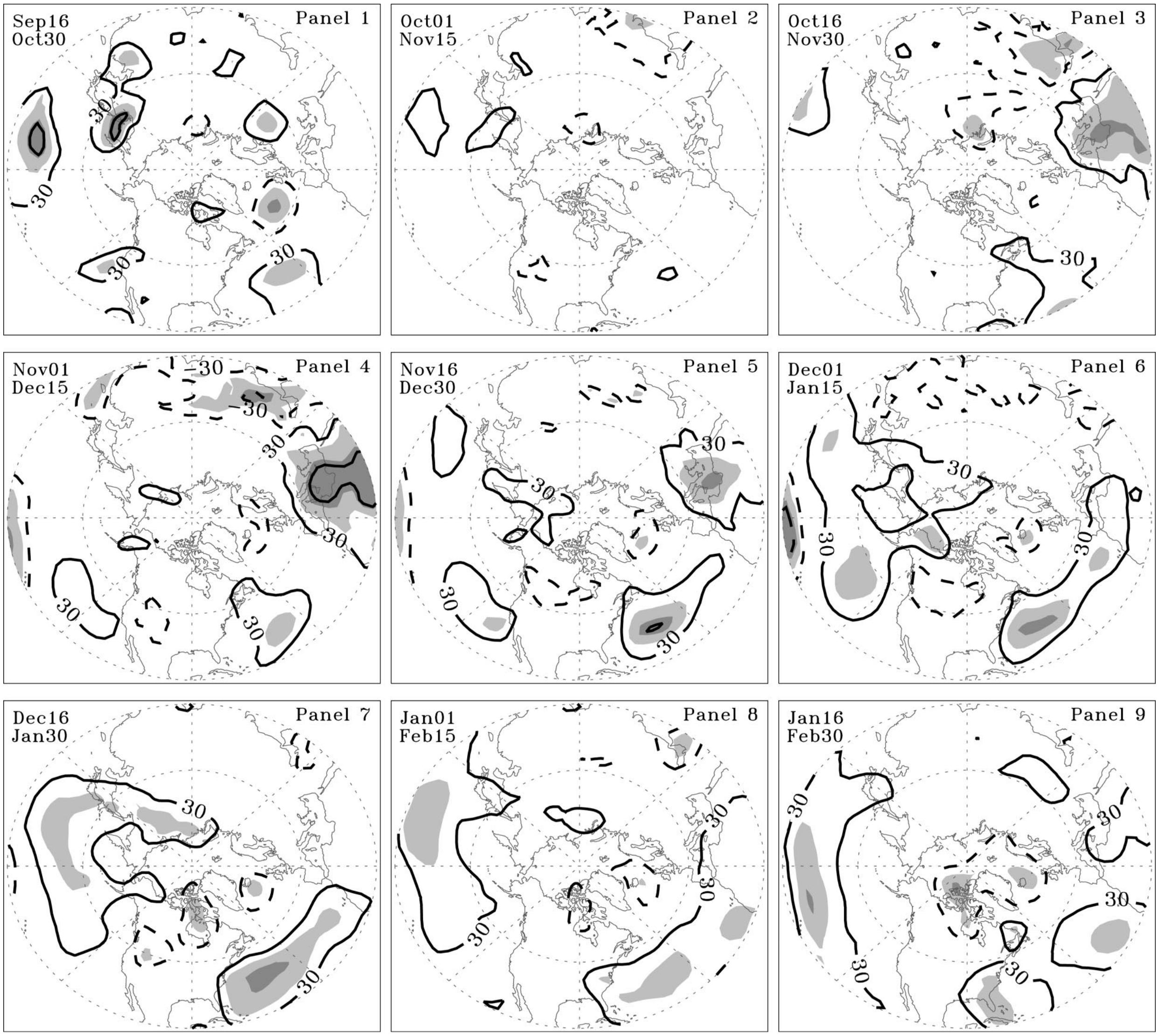

FIG. 8. Percent correlation between EOF1 expansion coefficients for DJF average Northern Hemisphere 50-hPa geopotential height field and a series of 45-day-average gridpoint SLP, for the FIX simulation. Contours drawn at ±30%, 60%, and 80% correlation. Solid (dashed) lines denote positive (negative) correlation. Light (dark) shading represents absolute correlations in excess of 44% (56%), representing 95% (99%) statistical significance.

cover area over Siberia during the FREE simulation (not shown). This appears to contradict previous observation-based analyses that do exhibit a statistically significant correlation between autumn Eurasian snow cover and winter climate indices (Cohen and Entekhabi 1999; Saito et al. 2001). Potential reasons for this discrepancy include the following: 1) although snow variations modulate the modeled AO–NAO, they are not the principal forcing mechanism (as indicated in section 3), so a direct correlation is not necessarily expected; and 2) internal snow cover and snowmass variability within the ECHAM3 model is notably less than that observed, so the magnitude of the snow forcing in the FREE simulation may be insufficient to yield a direct relationship with the AO–NAO pattern. For example, the mid-October snow-covered area over Siberia (defined as 1.6×10^7 km^2 of land surface area within 45°–90°N and 45°–135°E) ranges from roughly 8.5×10^6 to 1.3×10^7 km^2 over the 20 realizations of the FREE simulation. In contrast, 20 yr (1972–92) of visible satellite observations (Robinson et al. 1993) range from 3.0×10^6 to 1.5×10^7 km^2, a spread nearly 3 times as large as for the FREE simulation.

Note that the results presented in Figs. 6 and 7 indicate a clear correlation between autumn SLP anomalies over Siberia and winter climate indices, which implies that the connection between Siberian snow and the overlying SLP is lacking in the FREE simulation, and suggests that snow forcing in the FREE simulation may be insufficient. Ongoing research involves additional

GCM experiments in which weekly snow cover and snow mass is prescribed based on observations during extreme high and low autumn snow cover years, and the ensemble mean response to these snow perturbations is evaluated. These experiments will build upon the results presented here by analyzing the effect of snow anomalies that are larger than those contained in the FREE simulation, yet still realistic. Also, the hypothesized surface and vertical teleconnection pathways, for example, the expansion and migration of semipermanent surface sea level pressure cells, and the vertical propagation of stationary waves, will be investigated in greater detail.

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