On the Influence of Ocean-Atmosphere Interaction on the Arctic Oscillation in two General Circulation Models

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Abstract

The influence of ocean-atmosphere interaction on the wintertime Arctic Oscillation (AO) is investigated using a hierarchy of experiments made with two general circulation models (GCMs), ranging from climatologically-forced atmospheric to fully-coupled ocean-atmospheric GCMs with increasing greenhouse-gas concentrations.

Both GCMs reproduce well the AO spatial pattern, defined by the leading hemispheric mode of monthly sea level pressure or daily 700 hPa geopotential height, although the North Pacific pole is more pronounced compared to observations. Coupling is not found to influence this spatial pattern.

Power spectra are examined for evidence of ocean-atmospheric interaction in the form of spectral reddening or significant spectral peaks. No measurable influence is found. On interannual time scales, all the model AO spectra are approximately "white," with no clear evidence of any statistically significant spectral peaks in the coupled experiments. Greenhouse-gas induced changes in sea-level pressure are found to project onto the AO in one of the GCMs but not the other. On sub-seasonal time scales, the spectra are "red" in all the model configurations, but ocean-atmosphere interaction is not found to amplify the redness. Significant spectral peaks are found in the 15–25 day period range, consistent with observed spectra.

Daily histograms of the simulated AO indices are found to be negatively skewed. A Gaussian mixture model is used to estimate the probability density function of daily hemispheric height maps, and yields three circulation regimes in both the simulations and observed data. The uncoupled atmospheric GCM simulations exhibit AO-like regimes that acquire stronger wave-like characteristics in the coupled runs.

1 Introduction

The Arctic Oscillation (AO) emerges as the leading empirical mode of wintertime monthly sea-level pressure (SLP) over the northern hemisphere (NH) (Lorenz 1951; Kutzbach 1970; Trenberth and Paolino 1981; Wallace and Gutzler 1981; Thompson and Wallace 1998). Its spatial structure is characterized by anomalous SLP of one sign throughout the Arctic basin, with anomalies of the opposite sign centered over the Azores and, more weakly, over the North Pacific (Fig. 1f, below). The AO index—which is highly correlated with the North Atlantic Oscillation (NAO) index so that the two are arguably almost synonymous (Deser 2000)—has exhibited large decadal-scale near-cyclic variations over the past 30 years (see Fig. 2b, below), together with an upward trend. Through advection, the circulation pattern exerts a strong impact on wintertime surface temperature, precipitation and storminess over Europe (e.g. Hurrell 1995).

It is presently unknown whether this recent temporal evolution contains a deterministic component, that would hence be potentially predictable. In the context of the observed NAO record, which is at best about 175 years long, two studies of its statistical properties have argued that the index's recent behavior is not statistically significantly different from noise (Wunsch

1999; Stephenson et al. 2000).

If, on the other hand, an underlying deterministic near-decadal component does exist, interactions with the ocean are a leading contender for an explanation. Recent ensemble simulations with atmospheric general circulation models (GCMs) suggest that the NAO's decadal evolution can be partly reproduced—at least in phase if not in amplitude—by prescribing the observed global evolution of sea-surface temperatures (SST) and sea ice (Rodwell et al, 1999; Mehta et al. 1999). This type of atmospheric GCM experiments have been criticized as unphysical by Bretherton and Battisti (2000) because an SST evolution is prescribed, giving rise to anomalous midlatitude air-sea heat fluxes that are of the wrong sign compared to those that would be obtained in a coupled system; they also point out that there is not necessarily any implied predictability associated with the ocean, because midlatitude SSTs are largely a product of the atmosphere itself. These experiments do, nonetheless, indicate some sensitivity to the ocean, whether this be in the midlatitudes or tropics.

The aim of this paper is to examine the role of ocean-atmosphere interaction on the AO using coupled ocean-atmosphere GCMs. By comparing fully coupled simulations with those made with the atmospheric component of the models alone with climatological SSTs prescribed, the role of the ocean should be clear. If GCMs can simulate the AO adequately, multi-century simulations also have, in addition, the potential to determine the AO's power spectrum with higher statistical confidence than can be done with the relatively short observational record. The AO paradigm is chosen here, rather than the NAO one, so as to take a general hemispheric measure of variability, rather than a sectorial one, and because both the GCMs considered here do simulate a more pronounced AO pattern than in nature; thus the AO is arguably more relevant to the GCM simulations than to the observed record.

The AO, like other atmospheric teleconnection patterns, appears to to be an intrinsic mode of the atmosphere with a time scale of the order of a week (e.g., Feldstein 2000). Ocean-atmosphere interaction can potentially influence both its spatial structure and temporal evolution. Empirical orthogonal functions (EOFs) of monthly hemispheric SLP will be used to characterize spatial structure, as well as recurrent circulation regimes defined from the probability density function (PDF) of daily hemispheric geopotential height maps. The leading EOF of SLP has been shown to represent the AO well (Thompson and Wallace 1998). However, several different approaches can be taken to project climate variability onto a few "modes." The multiple-regimes approach is attractive because it involves no linearity assumption.

Power spectra of the AO's principal component time series will be used to examine temporal evolution. Two types of temporal changes may result from oceanic coupling. Coupling could potentially produce a spectral peak, either through "loop-type" oscillations (e.g. Latif and Barnett 1994) or through "spatial resonance" (e.g. Saravanan and McWilliams 1997). Coupling can also "redden" the shape of the spectrum by reducing the thermal damping of the atmosphere at the sea surface and thus increasing persistence of atmospheric anomalies. This reddening mechanism has been demonstrated in a simple stochastically-forced 1-D atmospheric energy-balance model coupled to an oceanic mixed layer (Barsugli and Battisti 1998).

Simulations are analyzed from a hierarchy of uncoupled-atmospheric and coupled ocean-atmospheric GCMs from two modeling centers; the models and simulations are presented in Section 2 and summarized in Table 1. We examine the spatial structure of simulated AO-like variability and its power spectrum, firstly using monthly SLP (Section 3), and secondly in terms of daily 700 hPa geopotential height fields (Section 4). A summary and discussion are given in Section 5.

2 The GCMs

Two different GCMs are considered. The first is the NCAR Climate System model (CSM), version one, which consists of the CCM3 atmospheric GCM and the NCOM oceanic GCM, run without flux correction (Boville and Gent 1998). Monthly data is available from this model. The second is the ECHAM/OPYC suite of models from the Max Planck Institute for Meteorology in Hamburg (Roeckner et al. 1999). Daily data is also readily available for this model, enabling the additional analysis of sub-seasonal time scales. Table 1 describes the models, and the set of simulations used. Comparisons against observed data are made using the NCEP/NCAR Reanalysis 1948–99 (Kalnay et al. 1996).

By using two different GCMs, we obtain some measure of the inter-model sensitivity of the results, and partially alleviate the difficulties associated with comparing the uncoupled and coupled models. In the case of the non-flux corrected CSM, some differences in climate between uncoupled and coupled simulations are inevitable, although these are generally quite small in the variables considered here (Boville and Hurrell 1998). In the flux-corrected Hamburg model, a more-recent version of the atmospheric model was used in the coupled simulations. The main differences between ECHAM3 and

ECHAM4 are a new radiation scheme and new closure for deep convection; both have minor impacts for our purposes (Roeckner et al. 1996).

All fields are for the northern hemisphere (NH), extend northward from 20°N, and were area-weighted by the square-root of cosine latitude. The seasonal cycle was removed at the outset by subtracting the ensemble mean over each calendar month or, in the intraseasonal case, by subtracting an average seasonal cycle, constructed by applying a 5-day running mean and averaging over all years.

3 Intermonthly and interannual variability

3.1 Spatial Structure

Figure 1 shows the leading EOF of monthly NH SLP for the November–April extended winter season. In all cases, EOF-1 exhibits a clear AO pattern (Thompson and Wallace 1998). It is well separated in variance from EOF-2 (see figure caption), which is characterized by a large anomaly over the Gulf of Alaska with weak wave-like features over North America and resembles the Pacific/North American (PNA) pattern; in the following the PNA/reverse-PNA terminology is used to refer to this kind of feature.

All the model simulations capture the basic observed tripole pattern rather well, and this is consistent with previous GCM studies (e.g. Lau 1981; Fyfe et al. 1999). The intensity of the North Pacific center appears to be exaggerated in both models, making the pattern more symmetric between the two ocean sectors than in observations. However, the observed AO patten derived by Thompson and Wallace (1998) using a century of data is more annular than the one derived here using the last 50 years alone (see Fig. 4c below). The robustness of the Pacific pole in the simulations was verified by splitting the data sets into two halves and repeating the analysis; it is also highly significant according to a Student t-test.

The impact of ocean-atmospheric coupling on the spatial pattern is minimal in both the Hamburg and NCAR models. Coupling slightly decreases the variance of EOF 1 in both models, while that of EOF 2 (a PNA-lke pattern) increases slightly (see figure caption).

The uncoupled AGCM simulations in Figs. 1a,b use seasonally-varying climatological SST distributions at the lower boundary. A CCM3 "GOGA" simulation forced by observed variations in SST 1870–1997 from the GISST dataset (Rayner et al. 1995) was also examined. Its leading EOF of SLP presents a very similar pattern (not shown) to its climatologically-forced counterpart in Fig. 1b. In this CCM3-GISST simulation, the leading two EOFs account for 25.6% and 11.6% of the SLP variance respectively. These numbers are closer to those of the coupled CSM rather than the uncoupled climatological run, consistent with the expectation that PNA-like variability (EOF 2 in all cases) is more pronounced when SSTs vary over the tropical Pacific.

Greenhouse gas (GHG) forcing has a noticeable impact in ECHAM4/OPYC, with the North Pacific center becoming even more pronounced, and the Azores center weakening. This contrasts with the CSM where the AO pattern is almost indistinguishable with or without GHGs; thus, the CSM's EOF-1 (Fig. 1d) is shown for the 300-year control simulation concatenated with the 130-year transient GHG run.

3.2 Spectrum of the observed AO

The power spectrum of the observed AO is plotted in Fig. 2a, in terms of a singular spectrum analysis (SSA: Vautard and Ghil 1989; Ghil et al. 2000) of the 1899–1997 AO index of Thompson and Wallace (2000), using December–February (DJF) averages. The SSA spectra are eigenvalue spectra of the autocorrelation matrix, with power (i.e. the variance accounted for by each eigenmode) plotted against the dominant associated frequency of the corresponding temporal EOF. A pair of eigenvalues with near-equal variance indicates the presence of an oscillatory component of the time series, i.e. a spectral peak.

Two oscillatory pairs of eigen-elements are identified at periods of about 9.0 and 2.4 years. Assessing statistical significance against a red noise null-hypothesis (Allen and Smith 1996), indicates the quasi-biennial mode to be significant at about the 95% level while the near-decadal mode is not highly significant. The 9.0-year component was found to be strongest in the DJF season shown. Figure 2b shows the AO time series together with reconstructed components (RCs) for each oscillatory pair. The 9.0-year RC pair is only pronounced since about 1960. Since Arctic data may be not be reliable prior to 1950, the analysis has been repeated using the NAO indicies of Hurrell (1995) and the Climatic Research Unit (CRU) of the University of East Anglia (Jones et al. 1997). The near-decadal component is recovered in both cases with a period of 8.0 years, and is found to be statistically significant at the 95% level against red noise in the longer (175 year) CRU series. Repeating the analysis with annual averages of the CRU series recovers a very similar statistically significant component with a period of 7.8 years. The existence of a near 8-year period over the North Atlantic sector has been known for some time (Rogers 1984); it is also present in North Atlantic SSTs (Moron et al. 1998), the Central England temperature record (Plaut et al. 1995), as well as over North America (Dettinger et al. 1995).

The 2.4-year component is close to the Nyquist frequency and may not be reliable. However, a quasi-biennial component of the AO has been found in the past (Trenberth and Paolino 1981), and Baldwin and Dunkerton (1999) suggest that the stratospheric QBO may tend to excite the AO.

3.3 Spectra of simulated AO indicies

Figure 3 shows SSA spectra for time series corresponding to the models' EOF-1 in Fig. 1, using DJF averages to match those plotted in Fig. 2. Very similar results are obtained with November–April averages or using the multi-taper method (not shown). All the spectra are "white" in appearance. For a quantitative measure, Table 2 gives the lag-1 autocorrelation coefficient γ of an AR(1) process fitted to each series, together with an estimate of γ computed from a robust estimate of the noise following Mann and Lees (1996). In the robust estimate, narrow-band spectral features are removed using a median filter, yielding a measure of the background spectrum; its effect can be clearly seen in the Hamburg GHG run, where the trend component is excluded.

Table 2 confirms that none of the GCM spectra differ significantly from white noise; thus coupling does not measurably redden the AO spectrum in either model. The spectrum of the CCM3-GISST run is also white, so that prescribing observed SST variations in an "AMIP-style" fashion does not measurably influence the AO's spectrum in CCM3. Only the Reanalysis data 1948–99 are significantly red, while Thompson and Wallace's (2000) longer observed record is white. Although the spectra in Fig. 3 are normalized, the variances and amplitudes of the EOFs in Fig. 1 indicate that there in no displacement of the entire spectrum toward higher power in the coupled cases. If anything, the variance of the AO deceases in the coupled runs.

Aside from the overall shape of the spectrum, SSA can identify oscillatory modes that may arise due to ocean-atmosphere interactions. Could the 9-year RC in the observed AO series in Fig. 2 be a result of such coupling? The spectra of the model simulations provide no supportive evidence for this conjecture. Although the CSM does show a weakly-significant peak near 12 years that is not present in the uncoupled CCM, the level of statistical significance is modest considering the 300-year length of the simulation. Thus, neither long coupled GCM exhibits a clear decadal or interdecadal component.

Both GCMs do show some marginally significant interannual components ($T < 5$ years). Since correlations with SST, plotted in Fig. 4, occur over the equatorial Pacific as well in the coupled models, these peaks may be associated with weak El Niño-Southern Oscillation (ENSO)-like variability. The GCM behavior appears not to be realistic in this respect, since ENSO and the AO are almost unrelated in observed data, as indicated by the absence of SST correlations over the equatorial Pacific in Fig. 4c. The correlations over the northern oceans in Figs. 4a–c are largely consistent with the AO atmospheric anomalies forcing the ocean (e.g. Cayan 1992). The observed correlations with SST over the North Pacific are perhaps surprisingly strong in Fig. 4c, given the weakness of the North Pacific center of the observed AO. However, this map was obtained using Thompson and Wallace's (2000) index (1904-94), which has a stronger Pacific pole than Fig. 1f.

3.4 Greenhouse gas induced trends

A recent positive trend is evident in the AO index and polar night jet (e.g. Graf et al. 1995; Hurrell 1995; Thompson and Wallace 1998), and an upward trend is simulated in some GCM GHG runs (Fyfe et al .1999; Shindell et al. 1999) but not others (P. Kushner, pers. comm., 1999). Despite the large degree of similarity in AO structure in the CSM and Hamburg models, the linear SLP trends predicted by the two GCMs under increasing GHG concentrations have contrasting spatial distributions, as illustrated in Fig. 5. These project very differently onto the AO, as demonstrated in Fig. 6 which shows how the linear trends project onto the leading 10 EOFs of the respective coupled control run.

The Hamburg model's trend in SLP clearly projects very strongly onto the AO. However, much of the SLP decreases over the northern continents in Fig. 5a are confined to the planetary boundary layer, so that the trend is less AO-like at 700hPa (not shown). In addition, at least part of the large AO projection in the Hamburg model appears to be due to changes over the North Pacific.

4 Intraseasonal variability

4.1 Spatial Structure

Atmospheric variability on the time scales of days to weeks has typically been examined at the 700 hPa level because of the longer availability of observed data there. Thus, we select this level for our analysis of intraseasonal variability. Daily 700 hPa data was only readily available from the Hamburg models, and due to the large volume of the daily data, the analysis of the long coupled runs is restricted primarily to the first 100 years of the ECHAM4/OPYC control run, and to the the last 100 years of the GHG run (i.e. the 21st century).

Figure 7 shows the leading EOF of unfiltered daily 700-hPa geopotential height data over the NH November–March, from the Reanalysis, and the Hamburg uncoupled, coupled, and GHG coupled runs. The pattern that emerges is similar to the AO one in SLP shown in Fig. 1, but is more annular. It is less well separated from EOF-2 in variance (see caption), which again resembles the PNA pattern in all cases. The midlatitude centers of EOF-1 are more zonally elongated and the Arctic pole extends less into the northern continents than in Fig. 1. By repeating the analysis using monthly means of

700 hPa height, these differences in pattern can be traced primarily to the use of daily as opposed to monthly-mean data, rather than to the different vertical levels.

Compared to the Reanalysis data, the Hamburg model captures the daily AO pattern very realistically, again with a tendency for the Pacific center to be too strong. As in monthly SLP, there is very little difference in pattern between the uncoupled (Fig. 7b) and coupled (Fig. 7c) models. Thus, to first order, ocean-atmosphere interaction has no impact on modal spatial structure. When increases in greenhouse gases are prescribed through the 21st century, the Atlantic pole becomes more confined to Europe.

4.2 Spectra

To examine the temporal spectra of the daily November–March EOFs, we construct multi-year continuous time series from the PCs by padding with zeros during the months April–October, and forming 5-day averages. Thus, each time series consists of 73 5-day averages per year. SSA is then applied to these 5-day series, and the spectra plotted in Fig. 8.

In contrast to the seasonal-mean spectra, these spectra have a strongly red character. This is also seen clearly in terms of the lag-1 autocorrelation values, given in the caption of Fig. 8. The shapes of the model spectra match the observed one quite closely. As in the case of seasonal means, there is no clear increase in spectral redness in the coupled runs, compared to the uncoupled one.

All the 5-day AO spectra—both simulated and observed—exhibit statistically significant spectral peaks (i.e. pairs of eigenvalues) in the 15–25-day period range. The models appear to capture these intraseasonal AO oscillations, which previous work suggests may be associated with mountain torque forcing (Lott et al. 2000). However, there are clearly some differences in the locations of the peaks (i.e. pairs of eigenvalues) and the overall shape of the spectra between the model simulations, over and above what would be expected from sampling variations, given that sub-monthly variability is very well sampled. The spectra do deviate from the AR(1) null-hypothesis at higher frequencies, especially in the GHG case.

4.3 Histograms

Histograms of the daily PC-1 time series are plotted in Fig. 9, (without 5-day averaging). All the distributions are moderately skewed toward the negative polarity of the AO, with a skewness that exceeds several times its own standard deviation (see figure caption).

On physical grounds and by analogy with stratospheric warmings, radiational constraints control the maximum coldness and depth that the polar vortex can attain, while it is possible to fill it in very efficiently by poleward advection of warmer air (see also Gillett et al. 2000). Similarly, a positive AO corresponds to an intensified midlatitude jet stream whose strength must be bounded due to surface drag. Previous studies (White 1980; Nakamura and Wallace 1991) have shown that skewness in local SLP tends to be positive on the poleward side of the mean jet stream position, and negative to the south of it. This is interpreted in terms of air mass cut-offs on either side of the jet. It is consistent with the negative skewness of the AO in Fig. 9, since the Arctic pole (negative SLP when the AO index is positive) dominates the AO tripole. Atmospheric skewness is also discussed by Holzer (1996). The tendency for model simulated modes to be skewed is also found in simplified models. In a study of zonal-jet vacillation in a simple primitive equations model, Koo et al. (2000) found very marked skewing of the same sense in the distribution of midlatitude jet intensity.

4.4 Gaussian mixtures

It has been argued that there is a close relationship between a system's underlying dynamics and its leading EOFs (Mo and Ghil 1987). On the other hand, the orthogonality constraint and assumed linearity often limits the dynamical scope of EOFs. EOFs 1 and 2 of the intraseasonal data are much less well separated in variance than in the case of monthly SLP, so that the dynamical significance of the EOF 1 is less clear. To test the robustness of the EOF results to the methodolgy, daily NH height maps were analyzed by fitting a Gaussian mixture model to the PDF constructed in the subspace of the leading two EOFs. This analysis is based on the weather- or circulation-regime paradigm, in which the PDF is assumed to be generated by essentially low-order (chaotic) dynamics, leading to preferred regions of the phase space, or "regimes" (Legras and Ghil 1985). Smyth et al. (1999) have shown that observed daily NH geopotential heights can be well described by a mixture-model, made up three Gaussian components. Cross-validated likelihood was used to determine the best value of k, the number of components or clusters.

Following the procedure of Smyth et al. (1999), we fit the mixture model in a 2-D subspace of the data defined by the two leading EOFs of low-pass filtered daily 700 hPa geopotential height. As before, the November–March daily data was first deseasonalized by subtracting the mean seasonal cycle of 5-day running means, prior to applying a 10-day low-pass filter; the resulting seasons are 128 days long starting on Nov 12. The *unfiltered* deseasonalized data were then projected into the EOF subspace of the filtered data, and the PDF analyzed.

The mixture model was computed for the 50-winter Reanalysis data set, as well as the 100-winter ECHAM3 and ECHAM4/OPYC control simulation data sets. For the GHG simulation, the entire 1860–2100 data set is used, detrended by considering only deviations from seasonal means.

Table 3 gives the cross-validated log-likelihood of particular values of k, the number of Gaussian components, together with the associated (posterior) probability of k, determined using cross-validation by repeatedly dividing the data at random 20 times into two equal subsets. The log-likelihoods are relative values and from which the posterior probabilities are derived via Bayes' rule, assuming that each value of $k = 1, ..6$ is equally likely a priori (Smyth et al. 1999). Corroborating the results obtained with 44 December–February observed winters by Smyth et al. (1999), $k = 3$ is clearly the most likely value of k in the Reanalysis data set. We also obtain $k = 3$ for the uncoupled ECHAM3 simulation and for the ECHAM4/OPYC control run.

For the ECHAM4/OPYC GHG run, the method indicates $k = 4$ to be most the probable.

The centroids of the Gaussians are plotted in physical space in Fig. 10. The regime centroids of the Reanalysis data have spatial patterns that resemble the three hemispheric regimes obtained by Cheng and Wallace (1993) and Smyth et al. (1999) very closely. These regime centroids show a greater degree of zonal asymmetry than the leading observed EOF in Fig. 7a, that was interpreted as the AO in Sect. 4.1 (the EOFs of the low-pass filtered data are very similar). The ECHAM3 simulation's regimes are more zonally symmetric and AO-like, with Regimes 1 and 2 resembling opposite polarities of EOF-1, with Regime 3 being close to the climatological mean. In the coupled control run, the influence of the PNA pattern becomes visible in Regimes 1 and (especially) 3, yielding patterns that are quite similar to the observed ones.

The regime centroids tend, in general, to fall off the axes in the 2-D EOF subspace and therefore are linear combinations of EOFs 1 and 2. Since EOF 2 is PNA-like in all cases, it appears that annular variability cannot be separated from wave-like PNA patterns, and this is especially true when ocean-atmosphere interaction is included. The regime centroids thus indicate that the AO is more complex than the leading EOF would indicate, in keeping with previous regime analyses (e.g. Kimoto and Ghil 1993; Corti et al. 1999; Monahan et al. 2000).

In the 1860–2100 GHG simulation, the mixture model suggests 4 regimes. However, the centroids of two of these are very similar in pattern, with the fourth regime being further from the origin. We thus show the result obtained with $k = 3$, which closely reproduces three of the four regimes. Of these three, Regimes 2 and 3 are quite close to those in the coupled control run, while Regime 1 is even more like the reverse PNA pattern, and is close to the axis of PC-2 (not shown).

The ECHAM4/OPYC GHG simulation's regimes (Fig. 10d) were computed from *intraseasonal* anomalies, after subtraction of the individual winter means. By construction, there can be no trend in regime membership. However, it is of interest to know how the GHG-induced trends project onto these strictly intraseasonal regimes, analogous to Fig. 6 for the SLP EOFs. Changes in the frequency-of-occurrence of circulation regimes has been found in GCMs with prescribed GHG increases (Monahan et al. 2000; Hsu and Zweirs 2000), while the observed record is more ambiguous (Corti et al. 1999; Hsu and Zweirs 2000). Figure 11 shows 30-winter running averages of the frequency of regime occurrence, using the unfiltered data. These were computed by (1) projecting the *unfiltered* 700-hPa height maps onto the leading two *intraseasonal* EOFs, and (2) associating each map's location in this 2-D PC plane with the closest regime centroid, using the intraseasonal regimes in Fig. 10d. The frequency-of-occurrence of Regime 1 shows an upward trend at the expense of Regimes 2 and 3 which both become less frequent. The trend is thus identified here with a reverse PNA-like pattern, in contrast to the AO pattern indicated by the EOF projection in Fig. 6.

5 Summary and Discussion

Both the Hamburg and NCAR GCMs reproduce well the spatial structure of the AO, defined as the leading EOF of monthly-mean SLP. The results clearly indicate the AO to be an intrinsic mode of the atmosphere, because it occurs in the uncoupled models forced by climatological SST. The AO also emerges as the leading EOF of *daily* December–March 700-hPa geopotential height maps in the Hamburg models, with rather minor structural differences compared to the monthly data.

Like the observed data analyzed by Smyth et al. (1999), and examined here using the NCEP/NCAR Reanalysis data set, the PDFs of the Hamburg models are found to be best approximated by *three* underlying Gaussian clusters, corresponding to recurrent circulation regimes. The spatial structures of the simulated regimes are found to be quite realistic, especially in the ECHAM4/OPYC coupled control run. However, larger differences in regime structure are found between the simulations, as compared to the leading EOF. While the regime patterns may be more prone to sampling variability, they do suggest that it may not be dynamically meaningful to separate the annular component from the PNA-like component. The latter is weak in the uncoupled ECHAM3, but becomes stronger in the coupled runs, consistent with the influence of ENSO-like variability.

The Pacific pole of EOF 1 is stronger than observed in the uncoupled and coupled runs of both GCMs. This is consistent with the intraseasonal regimes that are more AO-like in the simulations than in observations, with the exception of the ECHAM4/OPYC GHG run. While the concept of an important annular mode in observed northern hemisphere SLP is controversial (Deser 2000), the concept does appear to be more relevant to the GCM simulations. An annular mode is clearly the dominant one in aquaplanet GCM studies (e.g., Feldstein and Lee 1996), and an overly-annular mode was found also found in the GCM study of Bladé (1997). Its relative strength in the GCMs studied here suggests that these models may be lacking in the physics that regionalizes the annular mode over the North Atlantic in reality. The climatological stationary waves are well simulated by both models. One possibility is that forcing from the tropical Pacific strongly disrupts the annular mode over the North Pacific in the real world. Even in the models, the simple notion of symmetry between the opposite polarities of the AO starts to break down in coupled runs, where model ENSO-like variability, though weak, appears to lead to a more vigorous PNA component that influences regime structure.

An upward trend in the AO index does occur in the GHG run of the Hamburg model. However, it is shown that this is due largely to changes over the North Pacific and that the trend in SLP does not resemble the AO pattern. The reverse PNA-like regime becomes particularly pronounced in the 1860–2100 GHG run, and its frequency-of-occurrence shows an upward trend. This finding is consistent with the study of Timmermann et al. (1999), who found that ENSO variability in the ECHAM4/OPYC model becomes skewed toward stronger La Niña events; these would tend to generate the reverse-PNA-like pattern which projects in this case onto the AO pattern giving rise to an apparent trend in the latter.

While the hemispheric patterns derived here are not appropriate for detecting regional scale features, we note that Ulbrich and Christoph (1999) have reported a systematic shift of the NAO's northern variability center in the same ECHAM4/OPYC experiment, from a position close to the east coast of Greenland, where it is located in the control run, to the Norwegian sea.

The spectrum of the observed AO, constructed from Thompson and Wallace's (2000) near-century long series using SSA, exhibits a near-decadal oscillatory component with a nominal periods 9.0 years. The latter is recovered with 95% statistical significance from Jones et al.'s (1997) 175-year record of the NAO, with a nominal period of 8.0 years. No clear evidence of similar oscillations was found in the coupled GCM simulations, despite the longer (300 year) records available. The simulated AO spectra are "white" in appearance, and the lag-1 autocorrelation is not found to increase significantly in the coupled simulations compared to the uncoupled ones.

Even on the intraseasonal time scale there is no measurable reddening of the spectrum in the coupled control run. In principal, we would expect atmospheric anomalies to be less thermally damped when SSTs are allowed the freedom to adjust to surface fluxes and thus for the spectrum to be redder (Barsugli and Battisti 1998). However, the model simulations suggest that this effect is negligible for the hemispheric-scale deep equivalent barotropic structure of the AO. This result can be compared to that of Bladé (1997) who investigated the effect of coupling using a low-resolution atmospheric GCM coupled to an oceanic slab mixed layer in a perpetual January experiment. Bladé (1997) found a large increase in low-level thermal variance due to the coupling, but only a very small increase in the 500-hPa geopotential variance. Nonetheless, she found a slight reddening of the AO spectrum at 500 hPa.

Intraseasonal oscillations with periods near 20 days are found in the AO, consistent with Lott et al. (2000) who have presented evidence that mountain torques play an active role in these oscillations. Similar significant spectral peaks are also found in the model simulations. However, the GHG simulation shows less-clear evidence of these oscillations, suggesting that interactions with topography might be less pronounced in this case due to differences of the AO's spatial distibution.

The spectra imply that ocean-atmosphere interaction exerts little influence on the AO time series. This finding agrees with Saravanan's (1998) finding for the NAO in CCM3 in which prescribing the observed monthly evolution of SST did not increase the variance of the NAO. However, it contrasts with Robertson et al. (2000) who found that the UCLA AGCM's NAO became much more vigorous when the observed monthly evolution of SST was prescribed over the Atlantic, compared to a control run with monthly climatological SSTs prescribed. It also contrasts with Rodwell et al. (1999) and Mehta et al. (2000) who were able to reproduce the low-frequency component of the NAO over the past half-century, including the trend, using large ensembles of AGCM simulations with the observed evolution of global SSTs and sea ice prescribed. Further work is required to resolve these important inter-model differences in result.

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Table captions

Table 1: Details of the GCM simulations.

Table 2: Interannual lag-1 autocorrelation (γ) of an AR(1) process fit to PC-1 of SLP. The robust estimate is computed from an multi-taper method (MTM) estimate of the spectrum smoothed with a median filter of width 0.017 cy/day, so as to remove narrow-band peaks. The 95th percentile of γ computed from 1000 Gaussian white-noise segments is 0.084, and 0.173 for the raw and robust estimates respectively.

Table 3: The cross-validated log-likelihood of the Gaussian mixture model, with the posterior probability (zero to one) given below it, as a function of k.

Table 1:

Model	Simulation Years	Description
ECHAM3	1–100	Hamburg uncoupled AGCM with climatological SSTs.
CCM3	1–100	NCAR uncoupled AGCM with climatological SSTs.
ECHAM4/OPYC Control	1–300	Hamburg coupled atmosphere-ocean GCM with fixed GHGs (flux corrected).
ECHAM4/OPYC GHG	1860–2100	Hamburg coupled atmosphere-ocean GCM with increasing GHGs according to IPCC IS92a scenario (flux corrected).
CSM Control	1–300	NCAR coupled atmosphere-ocean GCM (CCM3/NCOM) with fixed GHGs.
CSM GHG	1–130	CCM3/NCOM with CO_2 increasing at 1%/yr starting in year 10.

Table 2:

Time series	Raw	Robust
NCEP/NCAR Reanalysis 1948-99	0.320	0.415
Thompson and Wallace 1899-1997	0.082	0.096
ECHAM3	-0.098	0.000
CCM3 Clim. SSTs	0.005	0.015
CCM3 GISST SSTs	-0.120	0.000
ECHAM4/OPYC Control	-0.022	0.065
ECHAM4/OPYC GHG	0.146	0.013
CSM Control	0.040	0.131
CSM GHG	-0.018	0.156

Table 3:

k	Reanalysis	ECHAM3	ECHAM4/OPYC Control	ECHAM4/OPYC GHG 1860–2100
1	-349.578	-343.880	-336.784	-343.042
	0.000	0.000	0.000	0.000
2	-348.701	-342.733	-334.826	-341.843
	0.002	0.000	0.000	0.000
3	-348.467	-342.394	-333.784	-341.704
	0.802	1.000	1.000	0.005
4	-348.523	-342.684	-334.236	-341.660
	0.000	0.000	0.000	0.952
5	-348.919	-342.556	-334.108	-341.686
	0.000	0.000	0.000	0.040
6	-348.910	-342.643	-334.159	-341.708
	0.000	0.000	0.000	0.003

Figure 1: Leading NH monthly EOF (November–April) of SLP. (a) ECHAM3 (uncoupled), (b) CCM3 (uncoupled), (c) ECHAM4/OPYC control, (d) CSM (control and GHG simulations concatenated together), (e), ECHAM4/OPYC GHG run, and (f) NCEP/NCAR Reanalysis. The contour interval is 1 hPa, and the zero and negative contours are dashed. The variance of the associated PC is given in brackets; PC-2 (not shown) accounts for 10.3, 9.7, 11.3, 12.2, 12.6 and 11.8% variance in the six cases respectively. In (a), SLP was derived from 1000 hPa geopotential height (Z) using the approximate relationship $P - 1000 = Z/8$ where Z is expressed in meters and P in hPa.

Figure 2: An SSA of the DJF-mean observed AO index of Thompson and Wallace (2000) for 1899/1900–1996/97. (a) SSA power spectrum using a window-width of $M = 30$ years, with normalized power in temporal EOF k on the ordinate, and the dominant associated frequency of temporal EOF k along the abscissa. The error bars denote the 95% confidence interval of a red-noise null-hypothesis, computed using a chi-squared test. (b) The original series with the two leading pairs of reconstructed components superposed; the variance and approximate period of each are given in brackets. The approximate spectral resolution is 1/30 cycles/year.

Figure 3: Interannual SSA power spectra of simulated wintertime mean AO indices, constructed using December–February averages of the monthly SLP PC-1. (a) ECHAM3, (b) CCM3, (c) ECHAM4/OPYC control, (d) CSM control, (e) ECHAM4/OPYC GHG, and (f) CSM GHG. The SSA window-width is $M = 30$ years, with details as in Fig. 2. In panels (d) and (f), the EOF in Fig. 1d is projected onto the respective CSM experiment.

Figure 4: Correlations between November–April means of the AO index and SST at each grid point, using PC-1 of (a) CSM control (b), ECHAM4/OPYC control, and (c) Thompson and Wallace's (2000) index with the 1904–94 GISST data set. The 95% significance threshold for the correlation coefficient in each case is (a) 0.11, (b) 0.13, and (c) 0.27 respectively. These values are computed from a 2-sided t-test with the number of effective degrees of freedom computed conservatively, following Davis (1976). Contour interval is 0.1.

Figure 5: Linear November–April SLP trends for (a) ECHAM4/OPYC (1960–2100), and (b) CSM GHG (year 10–109). Contour interval is 1 hPa/(100 yrs).

Figure 6: Projections of linear SLP trends in Fig. 5, north of 20°N, onto the leading 10 monthly SLP EOFs of the respective coupled control run. The ordinate is in percent of the total projection onto the 10 EOFs.

Figure 7: EOF-1 of daily hemispheric 700 hPa geopotential height (November–March). (a) NCEP/NCAR Reanalysis (1948–99), (b) ECHAM3, (c) ECHAM4/OPYC control, (d) ECHAM/OPYC GHG. Contour interval is 10 gpm, with zero and negative contours dashed. The variance of the associated PC is given in brackets in percent of total; PC-2 (not shown) accounts for 7.8, 8.4, 7.8, and 8.9% of the variance respectively.

Figure 8: Intraseasonal SSA power spectra of AO constructed from 5-day means of PC-1 of daily data (cycles/day). A window-width of 1 year (i.e. $M = 73$ 5-day means) was used, with details as in Fig. 2. (a) NCEP/NCAR Reanalysis (1948–99), (b) ECHAM3, (c) ECHAM4/OPYC control, (d) ECHAM4/OPYC GHG. Robust MTM estimates of γ computed over the frequency range 1/(4 months) - 1/(5 days) are (a) 0.585, (b) 0.606, (c) 0.578, and (d) 0.651.

Figure 9: Histograms of daily PC-1. (a) NCEP/NCAR Reanalysis (1948–99), (b) ECHAM3, (c) ECHAM4/OPYC control, (d) ECHAM4/OPYC GHG. The skewness $\mathrm{Skew}(x_1 \ldots x_N) = \frac{1}{N}\sum_{j=1}^{N}\left[\frac{x_j - \bar{x}}{\sigma}\right]^3$ is noted in each panel. Its standard deviation is given approximately by $\sigma_{skew} \simeq \sqrt{15/N}$ (e.g. Press et al. 1987), which yields $\sigma_{skew} \simeq 0.142$ in (a) and $\sigma_{skew} \simeq 0.102$ in panels (b)–(d), assuming $N/10$ effective degrees of freedom.

Figure 10: Centroids of Gaussian mixture model applied to low-pass filtered daily data. The maps show the 700 hPa height anomalies at each centroid, contour interval 10 gpm. (a) NCEP/NCAR Reanalysis (1948–99), (b) ECHAM3, (c) ECHAM4/OPYC control, (d) ECHAM4/OPYC GHG 1860–2100 (detrended).

Figure 11: Thirty-year running means of regime frequency-of-occurrence for the ECHAM4/OPYC GHG simulation.