Zonal-Eddy Dynamics of the North Atlantic Oscillation

ERIC DEWEAVER AND SUMANT NIGAM
Department of Meteorology, University of Maryland at College Park, College Park, Maryland

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ABSTRACT

This research is an attempt to understand the dynamical mechanisms that drive the wintertime North Atlantic oscillation (NAO) on monthly and longer timescales. In an earlier work by DeWeaver and Nigam, the authors showed that momentum fluxes from stationary waves play a large role in maintaining the zonal-mean zonal wind ($u$) perturbations associated with the NAO. In this paper, a linear stationary wave model is used to show that zonal-mean flow anomalies in turn play a large role in maintaining the NAO stationary waves. A strong two-way coupling thus exists between $u$ and the stationary waves, in which each is both a source of and a response to the other.

When forced by zonal-eddy coupling terms—terms that represent the interaction between NAO-covariant zonal-mean zonal wind anomalies and the climatological eddy flow—together with heating and transient fluxes, the model produces a realistic simulation of the observed stationary wave pattern. Zonal-eddy coupling terms make the largest contribution to the simulated stationary waves. Every feature of the stationary wave pattern is forced to some extent by zonal-eddy coupling, and the upper-level trough over Greenland is forced almost entirely by the coupling terms. The stationary waves generated by zonal-eddy coupling are well positioned to provide additional momentum to the $u$ anomalies, demonstrating the strong positive feedback between zonal-mean and eddy flow components.

The NAO is known for its effect on tropospheric temperatures over northern Eurasia, and the model produces a realistic simulation of these temperature changes at midtropospheric levels. Zonal-eddy coupling, including the zonal advection of land–sea thermal contrasts, is partly responsible for the temperature changes. However, diabatic heating anomalies associated with the displacement of the Atlantic storm track are also influential, causing more than half of the warming over Scandinavia and most of cooling from North Africa to the Caspian Sea.

1. Introduction

Expansions and contractions of the polar vortex exert a profound influence on the winter climate of the Northern Hemisphere. Such vortex fluctuations are evident in Thompson and Wallace’s (1998, 2000) work on the Arctic Oscillation (AO), Hurrell’s (1995a) study of the North Atlantic oscillation (NAO), and Ting et al.’s (1996) discussion of their zonal index. But despite the efforts of these and earlier authors (e.g., Wallace and Hsu 1985; Lorenz 1951; Namias 1950; Rossby and Willett 1948; Rossby et al. 1939), no specific cause for the fluctuations has ever been found. A variety of modeling studies (e.g., Ting and Lau 1993; Branstator 1992; Lee and Feldstein 1996; Robinson 1996; Yu and Hartmann 1993; James and James 1992) show that such behavior can arise spontaneously from internal atmospheric dynamics. The chaotic nature of unforced atmospheric motions would then suggest that these vortex fluctuations limit our ability to predict climate (e.g., Hoerling et al. 1995). On the other hand, low-frequency variations of the polar vortex could be influenced by external factors such as SSTs (Rodwell et al. 1999; Ting et al. 1996), snow cover (Cohen and Entekhabi 1999; Kodera and Koide 1997), and anthropogenic emissions (Perlwitz and Graf 1995). In that case, our ability to understand and predict climate change could depend on our knowledge of vortex fluctuations and their dynamics.

Of course, polar vortex expansion and contraction is not just a Northern Hemisphere phenomenon. Similar fluctuations have been documented in the Southern Hemisphere (e.g., Thompson and Wallace 2000; Hartmann and Lo 1998; Nigam 1990; Karoly 1990; Kidson 1986; Trenberth 1984) and in simple atmospheric models with no orography or land–sea contrasts (e.g., Robinson 1996; Lee and Feldstein 1996; Yu and Hartmann 1993; James and James 1992). To emphasize the similarity of all these fluctuations, Thompson and Wallace (2000), have coined the term “annular mode,” which refers collectively to all manifestations of this type of behavior. Yet fluctuations in the monthly mean size and strength of the wintertime North Polar vortex differ from
The prominence of the stationary waves is illustrated in Fig. 1. Figure 1a shows the 250-mb streamfunction anomalies associated with the NAO mode of DeWeaver and Nigam (2000), which was found to be representative of polar vortex expansion and contraction. While the streamfunction clearly shows the opposition between polar and off-polar latitudes common to annular modes, it also shows strong departures from zonal symmetry, which are plotted separately in Fig. 1b. Comparison of Figs. 1a and 1b shows that the eddy contribution to the flow anomalies is comparable to the zonal-mean contribution over eastern North America, the North Atlantic, and much of Eurasia—that is, over half of the northern extratropics. It would thus appear that any diagnosis of annular fluctuations in the Northern Hemisphere should include a careful examination of the factors responsible for producing the associated eddies. In particular, to what extent are the zonal-mean anomalies responsible for the existence of the eddies?

In this paper, we use a baroclinic diagnostic model to examine the mechanisms that maintain the eddy components of the monthly mean NAO mode during the northern winter. Using the model, we identify the contribution of the zonal-mean components in forcing the eddies, together with the contributions to eddy forcing by anomalous transient fluxes and diabatic heating. The diagnosis reveals that the zonal-mean components do indeed play a dominant role in forcing the eddies, accounting for more than half the eddy amplitude of the NAO mode.

The dominant role of the zonal-mean components in forcing the eddies is particularly intriguing in light of DeWeaver and Nigam’s (2000) complementary finding that the eddies of the NAO mode also play a central role in maintaining the zonal-mean components. Specifically, the interaction between the eddy components and the climatological stationary waves was found to provide the largest momentum source for the zonal-mean zonal wind \( \tau \) anomalies of the mode. The evidence of both papers taken together thus demonstrates the existence of a strong two-way coupling between the zonal-mean and eddy components of the mode, in which each is both a source of and a response to the other.

The idea of eddy circulation anomalies produced by zonal-mean flow changes has a long history, dating back to the classic paper of Rossby et al. (1939). Since then, many studies have shown how changes in \( \tau \) lead to changes in the propagation of stationary waves (e.g., Dickinson 1968; Hoskins and Karoly 1981; Held 1983; Nigam and Lindzen 1989). Such dispersion studies generally imply that large-amplitude eddy anomalies will occur during vacillations of the zonal-mean circulation even in the absence of anomalous external forcing. Of particular interest to the present case are the studies of Branstator (1984) and Ting et al. (1996), which show observational evidence that the \( \tau \) anomalies associated with the Northern Hemisphere zonal index are responsible for much of the concurrent eddy activity.

Like the present work, these studies use diagnostic models to determine the eddy response to zonal-mean flow anomalies. The effect of the zonal-mean anomalies is determined by examining the models’ response to zonal-eddy coupling terms, which express the eddy forcing produced by the interaction of the zonal-mean flow anomalies with the asymmetries of the climatology. For example, zonal-eddy coupling terms include the thermal forcing that occurs due to the advection of climatological land–sea contrasts by \( \tau \) anomalies.

Although these papers strongly suggest that zonal-eddy coupling will have some effect on the eddy components of the NAO mode, they do not guarantee that this coupling will be the most prominent driver of the eddies. In both studies, there are substantial differences between the response to zonal-eddy coupling and the observed eddy patterns, particularly in the North Atlantic sector, a key region in our analysis (compare Branstator’s Figs. 4a and 8a, or Figs. 2a and 6a from Ting et al.). They find instead that the Pacific sector is the region of greatest correspondence between the observed

![Fig. 1. The 250-mb streamfunction anomalies covariant with the NAO: (a) total anomalies, including zonal-mean and eddy components, and (b) eddy components. Contour interval is 10^6 m^2 s^{-1}, with dark (light) shading for positive (negative) values in excess of 10^6 m^2 s^{-1}.](image)
eddy and the eddies forced by zonal-eddy coupling. Ting et al. refer specifically to the similarity between their solution and the Pacific–North American (PNA) pattern of Wallace and Gutzler (1981). But our research (DeWeaver and Nigam 2000) shows that PNA variability is both spatially and dynamically distinct from the polar vortex fluctuations of interest here. Thus the implications of the above results for the dynamics of the NAO mode are not clear.

Alternatively, Hurrell’s (1995b) work suggests that the eddy components of the NAO could occur as a response of the monthly mean flow to zonal-vorticity fluxes even in the absence of zonal-eddy coupling. The north–south shift in the Atlantic storm track that accompanies the NAO is well known (e.g., Rogers 1990), and Hurrell’s Fig. 11 shows that on the monthly timescale the shift produces a strong anomalous upper-level vorticity source. Furthermore, the centers of streamfunction forcing associated with this source are largely collocated with the streamfunction anomalies themselves (compare his Figs. 11 and 9). Thus “the (synoptic) eddies generally reinforce and help to maintain the anomalous upper-tropospheric rotational flow for a high NAO index” (p. 2297). The implication of this statement is that the pattern of anomalies can be understood from the pattern of synoptic forcing alone, without reference to additional dynamical processes like zonal-eddy coupling. Similar results were also reported by Lau (1988) and Lau and Nath (1991), who examined the synoptic forcing of the upper-level height anomalies that accompany the leading modes of Atlantic storm track variability.

Zonal-eddy coupling, heating, and NAO temperature change

The NAO is best known for its effect on the severity of the northern European winter (see van Loon and Rogers 1978 for historical references). Hurrell (1995a) reports that, for the December to March season, an increase of one standard deviation in the traditional NAO index yields surface temperature increases in excess of 2 K from Scandinavia to western Siberia. Rodwell et al. (1999) claim a correlation of 0.82 between this index and wintertime (December–February) surface temperatures in northern Europe (50°–70°N, 5°–50°E). Given the substantial human impact of such temperature variations (e.g., Lamb 1995), a detailed account of their generation and maintenance is clearly desirable.

Thompson and Wallace (1998, 2000) explain the temperature perturbations associated with the AO, which has a correlation of 0.85 with our NAO mode, in terms of zonal-eddy coupling in the thermodynamic equation. Their Fig. 10 shows that the advection of the climatological temperature field by the AO-covariant \( \vec{\mathbf{v}} \) anomalies produces a pattern of thermal forcing resembling the lower-tropospheric temperature anomalies of the AO. The AO stationary waves can then be thought of as the “hydrostatic signature” of the temperatures forced by zonal-eddy coupling. This analysis builds on the traditional understanding that enhanced westerlies during positive NAO excursions bring mild winters to Europe by onshore advection of temperate maritime air, while at the same time bringing severe winters to Greenland and the Atlantic by offshore advection of cold Canadian air (van Loon and Rogers 1978). However, we find that advection by \( \vec{\mathbf{v}} \) only accounts for about half of the thermal advection anomalies.

We turn to the linear model for a more complete diagnosis of the European temperature anomalies. The model produces a realistic simulation of the midtropospheric temperature changes, which are quite similar to their low-level counterparts. In addition to the zonal-eddy coupling terms, Atlantic sector heating anomalies play an important role in forcing the simulated temperature anomalies. The meridional displacement of the Atlantic storm track results in a dipole heating anomaly, which produces warming over Scandinavia and cooling over the Mediterranean region. Of course, the claim that heating plays an important role can never be made with complete confidence, since the strength of the response depends on the strength of the heating, which cannot be determined precisely from the available observations. To support the claim of a strong response to heating, we have compared the NAO-covariant heating rates from two independent reanalyses, and examined the precipitation anomalies from our primary dataset and two observational precipitation datasets. The comparisons, presented in appendix A, show general agreement in the precipitation patterns, though there are significant disagreements regarding the strength of the anomalies.

The remainder of this paper is divided into 6 sections. Section 2 describes the data used in the study and the derivation of the NAO mode. Section 3 discusses the model, and section 4 shows the diagnostic simulation. Section 5 considers the feedback of the eddies onto the zonal-mean NAO components, and examines our attempts to model the coupling process interactively in a model that simulates the zonal-mean and eddy components together. Section 6 presents the diagnosis of midtropospheric temperature anomalies over Europe. Concluding remarks follow in section 7.

2. Data sources

a. Primary data source

The primary data source for this study is the reanalysis carried out by the National Centers for Environmental Prediction—the National Center for Atmospheric Research (NCEP–NCAR). The data, described by Kalnay et al. (1996), are available online from NCEP in monthly mean pressure level format, and we have used data for all winter months (December–February, or DJF) from January 1958 to February 1998. The monthly means are averages of four 6-hourly analyses for each
day, computed on a 2.5° × 2.5° grid. The actual resolution of the height, wind, temperature, and vertical velocity fields is somewhat less than the grid spacing, since NCEP truncates these fields spectrally during postprocessing at total wavenumber 36 (T36). To limit the size of the dataset, we reduced the longitudinal grid spacing from 2.5° × 5° by removing grid points corresponding to longitudes not divisible by 5.

The forcing of the monthly mean flow by submonthly momentum fluxes was calculated from the convergence of the velocity covariance tensor, \( -\nabla \cdot (VV') - \frac{\partial (\omega' V')}{\partial \phi} \). Here \( V' \) and \( \omega' \) are the departures of the 6-h horizontal velocity vector and pressure vertical velocity component from their monthly mean values, and the subscript \( i \) denotes the average for the \( i \)th month. The velocity covariances were taken from the monthly mean pressure level reanalysis, and their derivatives were calculated by centered differencing. Likewise, forcing by submonthly thermal transients was calculated from the convergence of the covariance terms, \( (V'T') \), and \( (\omega' T') \), where \( T' \) is the 6-h departure of temperature from its monthly average.

Thermal transients are also used to produce the diabatic heating field, which is calculated residually in each month from the monthly mean thermodynamic equation (Nigam et al. 2000). Horizontal winds and temperatures are available on 17 levels from 1000 to 10 mb, but the pressure vertical velocity and its covariances are only available on 12 levels, from 1000 to 100 mb. Consequently, forcing to the diagnostic model by heating and submonthly transients is interpolated to zero above 100 mb.

b. The NAO mode and NAO-covariant anomalies

The NAO mode used here is the leading mode of a rotated principle component (RPC) analysis of the monthly mean DJF height field at 200 mb. A complete description of the analysis can be found in DeWeaver and Nigam (2000). Our use of a zonally varying pattern to measure polar vortex expansion and contraction may seem unnecessary, since only zonal-mean anomalies change the overall size and strength of the vortex. However, as discussed in DeWeaver and Nigam, the zonally averaged flow can be influenced by regional fluctuations and wavelike teleconnections that do not constitute annular behavior. To identify coherent annular fluctuations, it is therefore appropriate to consider the full horizontal variation of the flow rather than the latitudinal variation of the zonal-mean components. It should also be noted that RPC analysis provides a somewhat stringent test for annularity, since VARIMAX rotation favors longitudinally localized patterns over hemispheric-scale structures like annular modes (Richman 1986).

Thompson and Wallace (1998) use the leading EOF of extratropical (20°–90°N) sea level pressure (the AO) to measure the expansion and contraction of the North Polar vortex. While the AO circulation anomalies are somewhat more zonally symmetric than those accompanying our NAO mode, the anomaly patterns are quite similar, and the temporal correlation between the two measures is 0.85. Thus it is not surprising that every figure in this study can be reproduced to a good approximation using the AO time series in place of the NAO time series. The same holds for the two-point NAO index of Hurrell (1995a), which has a temporal correlation of 0.95 with our NAO mode.

We identify the NAO-related anomalies in a field of interest (streamfunction, temperature, heating, etc.) by taking the inner product of the normalized NAO time series with the monthly mean anomaly of the field at each grid point. The resulting pattern represents the dimensional anomaly of the field that would occur, on average, for an NAO excursion of one standard deviation. Fields labeled as NAO-covariant, NAO-related, or simply NAO fields (e.g., NAO heating), are all derived in this way. In particular, eddy components of the NAO-covariant circulation are referred to as NAO-covariant stationary waves.

c. Residual heating diagnosis

Of all the data used in this study, the residually diagnosed heating anomalies are the most uncertain. Of course, if zonal-eddy coupling were the only source of eddy anomalies accompanying the NAO, uncertainties in the heating would be of little consequence. But our results suggest that NAO heating contributes substantially to eddy divergence anomalies in low- and mid-latitudes (Fig. 4b), as well as midtropospheric temperature anomalies over western Europe (Fig. 9d). While the quality of the heating cannot be established from the available data, we show in this section that the pattern of the anomalies matches our expectations of NAO-related heating. Comparisons with rainfall datasets and an additional residual heating field are provided in appendix A.

Figure 2a shows the column mean of the NAO heating. The most prominent feature is the meridional dipole in the North Atlantic, which is suggestive of changes in latent heating due to the movement of the Atlantic storm track. Storm track shifts occurring in association with the NAO have been documented by Hurrell (1995a) and Rogers (1990). In their analysis, the eastern end of the storm track shifts southward during the negative phase of the NAO, so that the storm track has a more zonal orientation, while the storm track develops a strong southwest–northeast tilt during the positive phase. The extension of negative heating anomalies onto the Iberian Peninsula is also consistent with Hurrell’s finding of reduced precipitation over southern Europe during the positive NAO phase.

Further insight into the Atlantic sector heating anomalies can be gained from the heating fields produced by the assimilating model used for the reanalysis. These fields are taken from short-term (6-h) forecasts which
Fig. 2. NAO-covariant DJF column heating. (a) Column mean of heating calculated as a residual in the monthly mean thermodynamic equation (Jan 1958–Feb 1998); (b) column heating taken from short-term forecasts produced as part of the reanalysis (Jan 1973–Feb 1996); and contributions to (b) from (c) latent heating and (d) vertical diffusion. Contour interval in all panels is 0.1 K day$^{-1}$. The column mean is a pressure weighted average taken from 1000 to 100 mb, excluding below ground data, and fields are smoothed using a nine-point smoother. Zero contours are suppressed and dark (light) shading is used for positive (negative) values in excess of 0.1 K day$^{-1}$.

use the reanalyzed fields as initial conditions. They were obtained from NCEP on a monthly mean basis for the period 1973–96, on a 192 x 94 Gaussian grid with 28 sigma levels. To match the residual heating data, they were interpolated to a 2.5° x 5° grid with 13 pressure levels from 1000 to 100 mb.

Since the heating fields are taken from the model, a complete breakdown of the heating into contributions by radiation, precipitation, and boundary layer diffusion is available. Of course, the model-generated heating is less observationally constrained than the residually diagnosed heating, since it comes directly from the model parameterization schemes, and will be influenced by any spinup or climate drift problems occurring over the forecast period. However, comparison of Figs. 2a and 2b shows a large degree of similarity between the model-generated and residually diagnosed heating fields, particularly in the North Atlantic sector.

While the dipole pattern in the North Atlantic is slightly weaker in the model-generated heating than in the residual diagnosis, the two estimates are in close agreement concerning the shape of the dipole, including the small secondary maxima over the Labrador Sea and off the northeast coast of Greenland. As expected, latent heating along the storm track (Fig. 2c) makes a large contribution to the model heating. The meridional dipole on the eastern side of Greenland in Fig. 2c is quite consistent with the storm track changes discussed by Hurrell (1995a) and Rogers (1990). But on the western side of the basin, the large positive anomalies extending along the southern coast of Greenland are due to thermal diffusion in the planetary boundary layer (PBL), shown in Fig. 2d. The maximum value of the PBL heating is nearly twice as large as the maximum for the storm track heating.

The large values of thermal diffusion in the western North Atlantic can be understood as a consequence of the stronger offshore winds that bring more cold Canadian air into contact with the relatively warm ocean waters during the positive phase of the NAO. The figure thus shows that, according to the NCEP data, PBL diffusion and thermal exchange with the underlying ocean play an important role in damping out the advective cooling that accompanies the NAO. Likewise, negative PBL heating anomalies are found over Scandinavia, where enhanced onshore winds bring maritime air into
contact with colder land surface temperatures during the positive NAO phase.

3. The diagnostic model

a. General description

Diabatic heating is just one of a number of factors, including zonal-mean flow changes and synoptic vorticity flux anomalies, which can generate NAO-related stationary waves. To determine how these factors combine to produce the observed stationary wave pattern, we simulate the zonally asymmetric flow anomalies using a steady-state spectral primitive equation model linearized about the full climatological flow. Since the model is linear, the simulated stationary waves can be decomposed into separate components produced by each of the forcing terms. As discussed in DeWeaver and Nigam (2000, section 3 and Figs. 10a,b), linearization results in the neglect of some steady nonlinear advection terms, but these were found to make negligible contributions to the momentum and thermodynamic energy budgets of the stationary waves.

The linearized dynamical equations are in the $\sigma$-coordinate form proposed by Bourke (1974), with the atmospheric state vector $S$ expressed in terms of vorticity, divergence, temperature, and the logarithm of surface pressure: $S = (\zeta, D, T, q)$. In the diagnostic simulation (Figs. 3 and 4), we solve for $S_a^e$, the NAO-covariant anomalous eddy flow, given the zonally varying DJF basic state $S_a = S_a^b + S_s$. Here $S_a$ and $S_s$ are the anomalous and climatological atmospheric states, and $S_a^e$ and $S_s^e$ are the zonal-mean and eddy components of $S$. The forcing used for the simulation consists of the zonal-mean basic state. One reason cited by Hoerling et al. for the use of the zonal-mean basic state is that the linearization of the dynamical operator to the zonal-mean components of the flow anomalies can be found in their appendix (see also Nigam 1994). The linearization of the $\sigma$-coordinate primitive equations is equivalent to equations (A5)–(A8) of Branstator (1990), with differences only in the dissipation terms, which are discussed below. The model is truncated rhomboidally at zonal wavenumber 15 and has 15 sigma levels (R15L15): 0.995, 0.975, 0.945, 0.915, 0.85, 0.75, 0.65, 0.55, 0.45, 0.35, 0.25, 0.175, 0.125, 0.075, and 0.025.

It must be emphasized that the steady-state model is not predicated on the assumption that NAO anomalies remain steady for periods of a month or more. Rather, it is predicated on the less restrictive assumption that the monthly mean-eddy anomalies occur in response to monthly mean forcing. Monthly mean anomalies could also reflect the behavior of unstable, optimal, resonant, or slowly decaying free modes present in the initial state for each month. The importance of free modes has been advocated in a variety of studies, including Simmons et al. (1983), Borges and Hartmann (1992), and Ferranti et al. (1990); although Borges and Sardeshmukh (1995), Sardeshmukh et al. (1997), and Newman et al. (1997) have argued strongly that atmospheric damping is sufficient to minimize the impact of such modes on timescales of 10 days or more.

Clearly, the evolution of free modes cannot be captured by subjecting a steady linear model to fixed external forcing. However, a stationary wave model only requires the eddies to be forced: if the eddies are forced by the zonal-mean anomalies, which are in turn forced by the eddies, the two together could still constitute a free mode. In that case, the total flow anomalies, including both zonal-mean and eddy components, could not be simulated as a steady forced response, but the stationary wave model would still succeed. This possibility is addressed in section 5b.

Our use of a stationary wave model linearized about a wavy basic-state climatology constitutes a departure from the earlier studies of Hoerling et al. (1995) and Ting et al. (1996), who performed their diagnoses using a zonal-mean basic state. One reason cited by Hoerling et al. for the use of the zonal-mean basic state is that the response to zonal-eddy coupling in this model can be interpreted using the zonal-mean refractive index arguments of Nigam and Lindzen (1989) and Kang (1990). While this sort of interpretation is quite desirable, we have found that the zonal-mean basic state produces a degraded simulation, although qualitatively similar results can be obtained.

b. Zonal-eddy coupling terms

Zonal-eddy coupling terms occur in all the primitive equations, and are formed by applying the linear dynamical operator to the zonal-mean components of the flow anomalies. Schematically, the flow is assumed to satisfy $L^e(S_a) = F_a^e$, where $L^e$ is the dynamical operator of the model. We write the operator as $L^e$ because the
model only solves for the eddies, so the linear operator calculates only the eddy tendencies produced by the interaction between the anomalous state vector and the climatology. To form an equation for $S^*$, we set $S_a = S^* + \bar{S}$ and use the linearity of $L^*$ to write $L^*(S^*) = F^* - L^*(\bar{S})$. The forcing due to zonal-eddy coupling is thus given by $-L^*(\bar{S})$.

As an example, consider the horizontal advection term $\mathbf{V} \cdot \nabla T$ in the thermodynamic equation, which yields $\mathbf{V} \cdot \nabla \bar{T} + \mathbf{V} \cdot \nabla T_a$ when linearized about the zonally varying climatology. In the stationary wave model, the linear terms are further decomposed into the terms $(\mathbf{V}_a \cdot \nabla \bar{T} + \mathbf{V}_a \cdot \nabla T_a^*)$ included in $L^*(S^*)$ and the zonal-eddy coupling terms $-\bar{\tau}_x \partial T^*/\partial x - \bar{\tau}_y \partial T^*/\partial y - \bar{\nu} \partial \bar{T}/\partial y$. The $\bar{\tau}_x$ term includes the cooling effect of the anomalous zonal—mean zonal wind advecting cold Canadian air into the Davis Strait/Labrador Sea region and mild marine air over northern Europe. Zonal-eddy coupling terms also represent the orographic uplift produced as $\bar{\tau}_x$ blows over Greenland (that is, the eddies in the climatological $q$ and $T$ fields caused by the Greenland orography), and the advection by $\bar{\nu}$ of the upper-level zonal vorticity gradients in the climatological jets. In addition to the advection terms, there are zonal-eddy terms in the vorticity and divergence equations derived from the solenoidal term $RT \nabla q$ ($R$ is the gas constant)

Fig. 3. (a) NAO-covariant eddy streamfunction at 0.25 $\sigma$. (b) Simulation from stationary wave model linearized about a zonally varying basic state. Contours and shading as in Fig. 1.
Fig. 4. (a) NAO-covariant eddy divergence at 0.25 s\(^{-1}\), and (b) eddy divergence simulation from stationary wave model. Contour interval is \(2 \times 10^{-7} \text{s}^{-1}\), with dark (light) shading for positive (negative) values in excess of \(2 \times 10^{-7} \text{s}^{-1}\), and zero contours suppressed. (c) NAO-covariant eddy surface pressure, and (d) eddy surface pressure from stationary wave model. Contour interval is 1 mb, with dark (light) shading for positive (negative) values in excess of 1 mb, and zero contour suppressed. The surface pressure anomaly \(p_a\) is calculated from the log(surface pressure) anomaly \(q_a\), using \(q_a = p_a/p_c\), where \(p_c\) is the climatological surface pressure.
in the $\sigma$-coordinate version of the pressure gradient force.

c. Dissipation

Without some form of dissipation, the matrix, which represents the linearized dynamical equations, is singular and diagnostic simulations cannot be obtained. Relatively large dissipation values are needed for models like ours that are linearized about the zonally varying, or wavy, climatology. For example, Ting and Lau (1993) use a biharmonic diffusion coefficient which is 10 times larger than the one used in the atmospheric general circulation model (AGCM) which they diagnose. Branstator (1992) claims that strong dissipation is required because the dynamical operator has numerous steady eigenmodes that would otherwise resonate with the imposed steady-state forcing. While the computational need for dissipation is clear, the physical significance of the large dissipation values is not. Recently, Ting and Yu (1998) presented evidence that dissipation serves as a proxy for nonlinear effects.

For the solutions shown here, dissipation takes the form of uniform horizontal and vertical diffusion together with a diffusive (Ekman) planetary boundary layer described in appendix B. Horizontal diffusion is applied in the vorticity, divergence, and temperature equations with a coefficient of $10^6$ m$^2$ s$^{-1}$, and is applied to the temperature field on isobaric surfaces using the approximation given in Hoskins and Karoly (1981). Following Branstator (1990, 1992), the vertical diffusion has the form $v \partial^2 \zeta / \partial \sigma^2$, where $\zeta$, $D$, and $T$ are the perturbation vorticity, divergence, and temperature fields; $\kappa = R/C_p$ ($R$ is the gas constant and $C_p$ is the specific heat at constant pressure); and $v = (1000 \text{ days})^{-1}$. Like Branstator (1990) and Ting and Hoerling (1993), we find that vertical diffusion is necessary to eliminate grid-scale structures in the vertical.

d. Solution method

We solve the linear equations for the model by Gaussian elimination, a process that involves some technical challenges due to the extreme size of the matrix involved (Branstator 1992; Ting and Lau 1993). To construct the matrix we use the tendency method of Hoskins and Karoly (1981), in which the linear dynamical operator is applied to a set of basis elements used to represent the atmospheric state vector. The matrix has the dimension of the state vector, so that a modest R15L15 resolution results in a roughly 22,000-dimensional matrix, occupying nearly 500 MW of memory. Such large amounts of memory are not currently available for single applications on conventional supercomputers. If, as in our case, the model is linearized about the full threedimensionally varying climatological flow, the resulting matrix is not sparse and has no known symmetries or other nice properties. Ting and Lau (1993) avoid the full matrix by casting the dynamical equations in the form proposed by Schneider (1989), in which the system is decoupled in the vertical. While this formulation results in a tractable problem, the strategy is difficult to implement (Ting and Held 1990; appendix A), and resolution is still somewhat limited.

Recently, parallel computers have become available that accommodate large matrices by spreading them over many processors, and our model is designed to exploit this new technology. Using 32 processors, the model can be solved on a Cray T3E at R15L15 resolution in about 15 min of wall clock time. The diagnostic model is relatively easy to implement as a parallel code, because the task of matrix generation can be assigned to multiple processors simply by giving each processor a subset of the basis elements. The tendency method can then be applied independently on each processor to produce a portion of the matrix, and the resulting distributed matrix can be inverted using standard parallel subroutines. Unlike Schneider’s (1989) method, no specific form of the linearized dynamical equations is required. The matrix is constructed using the Message Passing Interface (Snir et al. 1996) in the block-cyclic form required by the Scalable Linear Algebra Package (Blackford et al. 1997), which is used to obtain solutions.

4. Diagnostic stationary wave simulation

The NAO-covariant stationary waves are represented by the eddy streamfunction field at $\sigma = 0.25$, shown in Fig. 3a. The $0.25\sigma$ level is chosen for consistency with DeWeaver and Nigam (2000, Fig. 9b), where the 250-mb level is studied because it is near the level at which the stationary waves provide the strongest forcing to the zonal-mean flow. The corresponding stationary wave simulation (Fig. 3b) captures most of the features in 3a, including the trough over Greenland, the ridge centered over the Baltic Sea, the low over the Caspian Sea, and the high over the northern Arabian Sea. The model is quite successful in simulating the longitudinal placement of all of these features, and the area-weighted correlation between the patterns in Figs. 3a and 3b is 0.8.

Of course, the simulation also has its shortcomings, like the exaggerated subtropical response near the date line. The chief deficiency of the simulation is the weakness of the model response: the Greenland low is underestimated by about one-third, and the Baltic high is weak by about 25%. The subtropical and midlatitude features in the Atlantic sector have less than 50% of the amplitude of their counterparts in Fig. 3a, and the Caspian low is equally weak. At high latitudes, the weakness of the upper-level stationary waves is partly a hydrostatic consequence of a weak temperature response in the boundary layer, where thermal forcing is dissipated by PBL diffusion. The search for better dissipation strategies, which produce a stable calculation while al-
lowing for a robust low level response, is part of our continuing research effort.

It could be argued that the streamfunction is the easiest target for a diagnostic simulation, since it emphasizes the largest scales. To show the performance of the diagnostic model on smaller scales, we display the 0.25σ NAO covariant and simulated eddy divergence in Fig. 4. Consistent with the NCEP heating and precipitation fields (Figs. 2a and A1b), the NAO-covariant divergence (Fig. 4a) has positive anomalies accompanying the tropical convection over northern South America and the subtropical North Atlantic. Upper-level convergence can also be found over the precipitation deficient regions in the midlatitude Atlantic and the Mediterranean. Some divergent outflow occurs over the region of enhanced rainfall in the northern North Atlantic, although the pattern is complicated by the presence of the Greenland orography, and the fact that high-latitude heating is balanced by horizontal advection as well as vertical motions. Most of these features are present in the simulated divergence in Fig. 4b, which has an area-weighted correlation of 0.77 with Fig. 4a.

Finally, we compare the simulated and observed patterns of surface pressure in Figs. 4c and 4d. The covariant pattern in Fig. 4c is somewhat different from the typical NAO sea level pressure pattern (e.g., Hurrell 1995a, Fig. 1b) because the zonal mean has been removed for comparison with the model simulation. Again, the correlation between simulation and observations is quite high at 0.9, and in this case the model simulates both the pattern and its amplitude reasonably well.

5. Response to zonal-eddy coupling

a. Response in the stationary wave model

Having produced a credible simulation, we can now ask how much of the stationary wave pattern is a response to the changes in the zonal-mean flow. Figure 5a shows that zonal-eddy coupling makes some contribution to every feature of the simulation (Fig. 4b). It accounts for nearly all of the amplitude of the Greenland trough, and much of the amplitude of the high over northern Siberia. The figure thus establishes the central conclusion of this paper, that zonal-eddy coupling plays the dominant role in maintaining the NAO stationary waves. It must be emphasized, though, that the stationary waves are not just a passive response to the changes in the zonal-mean flow, for stationary waves play a very active role in maintaining the zonal-mean anomalies.

Figure 5 of DeWeaver and Nigam (2000) shows how π anomalies can be maintained by the interaction between climatological and anomalous stationary waves. In that figure, a large portion of the momentum flux is generated in the Atlantic sector, the region that has the strongest stationary wave response to zonal-eddy coupling. Thus it appears that when the NAO-covariant π pattern is excited for any reason, it automatically generates stationary waves through the zonal-eddy coupling mechanism, which act to further enhance and sustain it. To illustrate this feedback, we have calculated the zonal-mean zonal momentum source produced by the stationary wave response to zonal-eddy coupling. The forcing to the zonal-mean zonal momentum equation due to the convergence of momentum flux from the interaction of anomalous and climatological stationary waves is given by \( \vec{F}_s = \vec{e}_y^* v^* + \vec{e}_y v^* \). Here \( v^* \) is the eddy component of the meridional velocity.

Figures 5b,c,d show the \( \vec{F}_s \) profiles that result when \( \vec{e}_y \) and \( v \) are taken from the observations, the model simulation, and the response to zonal-eddy coupling. The observed \( \vec{F}_s \) in c is equivalent to the stationary wave forcing in Fig. 9b of DeWeaver and Nigam (2000), which was found to be the most prominent momentum source for NAO-covariant anomalies. Consistent with the overall weakness of the simulation, the forcing to \( \vec{F}_s \) by the simulated stationary waves in c is about 15% weaker than the forcing by the observed stationary waves. Figure 5d shows that the response to zonal-eddy coupling accounts for more than half of the total forcing by the simulated stationary waves. It thus appears that to a large extent, the \( \vec{F}_s \) anomalies are forced by the stationary waves, which they themselves maintain through zonal-eddy coupling.

b. Interactive modeling of the zonal-eddy feedback

Like Branstator (1984) and Ting et al. (1996), we have assessed the contribution of zonal-eddy coupling to stationary wave maintenance using a diagnostic model that solves only for the eddy flow components. The eddy-only model suffices to reveal the importance of zonal-eddy coupling in comparison to the direct forcing of eddies by transients and heating, but it can only represent zonal-eddy coupling as an external forcing to the stationary waves. If the zonal-mean anomalies are largely maintained by the stationary waves, which are maintained in turn by a combination of zonal-eddy coupling and external forcing, only a model that represents zonal-eddy coupling as an interactive, two-way process can give a complete account of the effects of the external forcing. For example, consider the scenario in which anomalous vorticity fluxes along the Atlantic storm track generate stationary waves, which interact with the climatological stationary waves to produce \( \vec{F}_s \) anomalies, which generate additional stationary waves, which further modify \( \vec{F}_s \) until an adjusted state is achieved. In an eddy-only model, zonal-eddy coupling appears as an independent forcing term, and there is no way to determine the contribution of the Atlantic transients to the \( \vec{F}_s \) anomalies.

On the other hand, if the diagnostic model could be solved for the full circulation anomaly, including the zonal-mean flow as well as the eddies, the response to transients would automatically include the effects of
zonal-eddy adjustment. Clearly, there are advantages to modeling the zonal-mean and eddy flow components together as a full circulation anomaly, and full anomaly models have been used in a number of diagnostic studies (e.g., Branstator 1990, 1992; Trenberth and Branstator 1992; Ting and Peng 1995; Ting and Hoerling 1993; Ting and Lau 1993; Navarra and Miyakoda 1988). From the success of these simulations one might conclude that zonal-eddy adjustment can be reasonably approximated by the linear dynamical operator of a diagnostic model. But despite the success of these models in a variety of situations, we have found that the zonal-eddy coupling associated with the NAO mode is not well captured by the dynamical operator of our model.

It is a simple matter to reconfigure the diagnostic model so that it solves for the zonal-mean anomalies as well as the stationary waves. The best full model solutions were obtained using Rayleigh damping at (20 days)$^{-1}$ rather than harmonic diffusion to dissipate the zonal-mean components, but results were qualitatively similar for both forms of damping. Damping is particularly important for the zonally symmetric anomalies, since the strong advection terms $\overline{u \partial(T_a, \zeta_a)} \overline{\partial x}$ are absent in the zonal-mean equations, and the planetary vor-
ticity advection ($\beta v_1$) is weak. These are the dominant terms through which the eddy circulations balance the imposed forcing. In the simple case of a zonally symmetric basic state with $\bar{\tau}_c = 0$, a balanced zonal-mean perturbation is a valid free solution to the undissipated equations: dissipation is therefore required to limit the strength of the steady forced zonal-mean response. Of course, such resonance can also occur for any nonzero zonal wavenumber for a given $\bar{\tau}_c$, but the resonance in the zonal-mean occurs for all $\bar{\tau}_c$.

The 0.25$s$ stationary waves from the full model simulation are shown in Fig. 6. For the tropical and midlatitude stationary waves, the full model simulation is quite realistic, and in some cases even better than the eddy-only simulation (Fig. 3b). The features over the southeastern United States and the tropical to midlatitude Atlantic that were underestimated in the eddy-only model are somewhat stronger, as is the connection between the lows off the African coast and over the Caspian region. But the high-latitude features that are most prominent in the response to zonal-eddy coupling (Fig. 5a), the Greenland trough and the northern Siberian ridge, are poorly simulated. The Greenland trough is centered over Baffin Island, well to the west of its observed position, and is considerably weaker than the Greenland trough in the eddy-only simulation. The northern Siberian ridge is largely absent, except for an erroneous center over eastern Siberia north of Kamchatka.

A poor simulation of the zonal-mean anomalies would be consistent with a poor simulation of the high-latitude stationary waves. If the zonal-mean flow anomalies are weak, their interaction with the climatological stationary waves will be weak, leading to a weak stationary wave response to zonal-eddy coupling. On the other hand, if the simulated stationary waves are weak, the zonal-mean zonal momentum source ($\bar{F}_u$) arising from anomalous-climatological stationary wave interactions will be weak, leading to a weak $\bar{u}$ response.

The $\bar{u}$ response from the full anomaly model is examined in Fig. 7. Figure 7a shows the NAO-covariant $\bar{\tau}$ anomalies, and Fig. 7b shows the $\bar{u}$ simulation from a truncated version of the diagnostic model that simulates only the zonal-mean flow anomalies. The zonal-mean only model is forced with the observed stationary wave momentum flux $\bar{F}_u$ as well as the zonal-mean components of the transient forcing and diabatic heating. In spite of its imperfections, the simulation from the zonal-mean only model shows a dipole structure similar to the observed $\bar{\tau}_a$, with an amplitude that is somewhat weak but generally comparable to the observations. Figure 7c shows the $\bar{u}$ simulation from the full model. As suggested by the weak stationary waves of the preceding figure, the northern cell of the $\bar{u}$ dipole is substantially weaker than its counterparts in the observations and the simulation forced by the observed $\bar{\tau}_a$. The forcing to $\bar{u}$ by the stationary waves from the full model simulation (Fig. 7d) is weaker than the forcing by observed stationary waves (Fig. 5b) by a factor of 2.

Apparently, the full model has difficulty simulating the mutual adjustment of the zonal-mean and eddy flow anomalies. We suspect that this difficulty has not been reported by previous authors because they applied their models to situations in which the coupling is weaker. Hoerling et al. (1995) claim, for example, that zonal-eddy coupling plays a minor role in establishing the stationary wave anomalies associated with the canonical El Niño. Ting and Lau (1993) diagnose an NAO-like...
Fig. 7. (a) NAO-covariant $\pi$ anomalies from NCEP reanalysis. (b) The $\pi$ simulation by a version of the diagnostic model that simulates the zonal-mean anomalies given the zonal-mean forcing by heating and transients, together with the zonal-mean forcing due to the interaction of the eddies in the climatological basic state with the observed NAO-covariant eddy flow anomalies. (c) The $\pi$ from the diagnostic simulation in Fig. 6. In (a)–(c), contour interval is 0.5 m s$^{-1}$, with dark (light) shading for positive (negative) values in excess of 0.5 m s$^{-1}$. (d) Forcing term $\mathcal{F}_a$ as defined in Fig. 5b, with $\zeta_a$ and $v_a$ taken from simulation in Fig. 6. Contours and shading for (d) are the same as in Fig. 5b.
pattern in AGCM output, and find that their diagnostic model produces an accurate representation of the stationary waves but a poor zonal-mean simulation. This behavior suggests that the two are not strongly coupled. They also state that the North American trough is weak in the AGCM climatology, and a weaker North American trough should lead to weaker interactions between the climatological eddies and the Atlantic sector flow anomalies.

One can imagine an extreme case in which the coupling becomes so strong that the zonal-mean and eddy anomalies maintain each other exclusively through their mutual interaction. An idealized example of this behavior is provided by the analytical model of Charney and Devore (1979; see also Held 1983), which in its simplest configuration possesses two stable steady solutions for the same external forcing. Such self-sustaining anomalies would not be diagnosable by a full anomaly model, since the model can only calculate the linear response to anomalous external forcing, and in this case the anomalies are not externally forced. It is thus reasonable to expect that full anomaly diagnosis will become more and more difficult as the strength of the coupling increases, and external forcing becomes a smaller and smaller part of the budget.

6. Diagnosis of temperature changes over Europe

Our diagnosis concludes with an examination of the most widely recognized climatic effect of the NAO: the temperature change over Northern Europe. It is of course the temperature change at the surface that matters most for human activities, and, as discussed in section 5a, the temperature anomalies at the lowest levels are rather weak in the diagnostic simulation. But while the NAO temperature pattern is strongest at the ground, it extends throughout the troposphere, and we have diagnosed the pattern at midtropospheric levels. The diagnosis thus offers insight into the warming of the land to the extent that the pattern of temperature anomalies has the same origins at the surface as it does aloft.

The preceding discussion suggests that zonal-eddy coupling should play an important role in bringing about the European temperature anomalies, and this is indeed the case. However, the stationary wave anomalies are not entirely explained by zonal-eddy coupling. In the observations and diagnostic simulation (Fig. 3), the positive streamfunction anomalies over northern Eurasia reach their maximum over the Baltic Sea, while the response to zonal-eddy coupling (Fig. 5a) is centered over northern Siberia, well to the east. Eddy forcing by heating and transients could thus play a role in establishing the circulation changes in the region around Scandinavia without the intervention of zonal-eddy coupling.

a. Observed 1000–500-mb thickness and advection anomalies

Thompson and Wallace (2000) propose that the lower-tropospheric temperature changes are forced by zonal-eddy coupling in the thermodynamic equation. As evidence, they cite the resemblance between the 1000–500-mb thickness anomalies and the temperature tendency induced in that layer by the term $-\overline{u_\theta\partial T^a/\partial x}$, the advection of the climatological temperature field by the anomalous zonal-mean zonal wind. The resemblance, while convincing, does not mean that zonal-eddy coupling in the thermodynamic equation is the only factor that affects the temperatures. To demonstrate that other factors are involved, we have plotted the complete horizontal temperature advection term $-(V\cdot\nabla T^a + \nabla \cdot \nabla T^a)$ in Fig. 8b, and the zonal-eddy term $-\overline{u_\theta\partial T^a/\partial x}$ in Fig. 8c, together with the covariant 1000–500-mb thickness (Fig. 8a). It should be noted that while the thickness anomalies in panel a are generally quite consistent with AO-covariant thickness in Thompson and Wallace (2000; Fig. 11), the AO pattern has a center over western Siberia that is stronger than the center over Scandinavia.

Both advection terms bear some resemblance to the thickness pattern, but the opposition between Canadian cooling and northern European warming is perhaps twice as strong in the total advection as it is in the zonal-eddy term. Other features of the thickness pattern also appear in the total advection term, like the Mediterranean cooling and the warming centered on the east coast of North America. The similarity of the thickness and horizontal advection patterns supports the view that much of the temperature pattern is due to onshore and offshore advection, though advection by $\overline{u_\theta}$ is only one component of that advection. Zonal eddy coupling terms in the other dynamical equations (e.g., $-\overline{u_\theta\partial T^a/\partial x}$) may also make indirect contributions by generating the eddy wind and temperature anomalies that contribute to the total thermal advection.

b. Simulation of 0.75σ–0.25σ thickness anomalies

To diagnose the midtropospheric temperature anomalies, we examine the NAO-covariant 0.75σ–0.25σ thickness in Fig. 9. Figure 9a gives the eddy thickness from the reanalysis, while Fig. 9b gives the corresponding solution from the eddy-only model. There is good
agreement between the midtropospheric pattern in Fig. 9a and the lower-tropospheric pattern of Fig. 8a despite the difference in altitude and the fact that we have removed the zonal-mean in Fig. 9a. The four-celled wave pattern extending from the Labrador Sea to the Arabian Sea is well represented in the simulation, although the lows are somewhat weak, and the high over northern Europe is overestimated by about 25%. The area-weighted correlation between the observed and simulated patterns is 0.87.

Figure 9c gives the response to the zonal-eddy coupling terms in all equations. The response shows a high-latitude dipole with strong cooling over Greenland and warming over northern Siberia. Zonal-eddy coupling produces most of the Greenland cooling. The fact that $\overline{\nu} \cdot \nabla T^* / \partial x$ is only about half of the thermal advection in the region indicates that the response to the coupling terms in the vorticity and surface pressure equations (not shown) is comparable to the response to the $\overline{\nu}$ advection term in the thermodynamic equation. It should be noted that the continental warming in the observations and simulation is centered over Scandinavia, while the response in Fig. 9c is centered well to the east over the arctic coast of Siberia.

The response to heating is plotted in Fig. 9d. It accounts for roughly half of the positive midtropospheric eddy temperature anomaly over Scandinavia, and much of the cooling to the south over the Mediterranean region. The diabatic heating used in the simulation (Fig. 9e) shows the meridional dipole associated with the displacement of the Atlantic storm track. It differs from the covariant heating in Fig. 2a because it has been truncated and interpolated to R15L15 resolution, and boundary layer heating has been removed (hence the lack of heating over the Labrador Sea). The temperature response downstream of the forcing is reminiscent of the propagation of stationary Rossby waves on a zonally symmetric climatological flow (Hoskins and Karoly 1981), despite the use of a wavy basic state. Figure 9f shows the thickness response when a zonal-mean basic state is used in the diagnostic model. The fact that the zonal-mean basic state yields such a similar response indicates that the temperature response can be understood in terms of classical stationary wave theory. Also, the cooling over the Black Sea is maintained largely by the local precipitation anomaly rather than by wave propagation from the Atlantic.

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Fig. 8. (a) 500–1000-mb thickness covariant with the NAO mode. Contour interval is 10 gpm, with dark (light) shading for positive (negative) values in excess of 10 gpm. (b) Horizontal thermal advection, given by $-\nabla \cdot \nabla T - \nabla \cdot \nabla T$, where $\mathbf{V}$ is the vector wind and $T$ is the temperature, averaged from 1000 to 500 mb. (c) Thermal advection by covariant $\overline{\nu}$, given by $-\overline{\nu} \cdot \nabla T^* / \partial x$, averaged from 1000 to 500 mb. In (b) and (c), contour interval is 0.2 K day$^{-1}$, with dark (light) shading for positive (negative) values in excess of 0.2 K day$^{-1}$, and zero contours suppressed.
Fig. 9. Diagnosis of eddy thickness anomalies (geopotential height differences) between the 0.25σ and 0.75σ levels. (a) NAO-covariant eddy thickness, (b) eddy thickness from stationary wave simulation in Fig. 3b, (c) eddy thickness from response to zonal-eddy coupling terms in Fig. 5a, (d) eddy thickness from response to heating alone. (e) Column heating used to produce thickness response in (d), and (f) response to heating when the stationary wave model is linearized about a zonal-mean basic state. In (a)–(d) and (e), contour interval is
Strictly speaking, the conclusion that storm track heating is an important cause of western European temperature fluctuations applies only to the NAO-covariant anomalies in the NCEP reanalysis. Some caution is in order when applying this conclusion to the real atmosphere, since it depends entirely on the strength of the heating anomalies in the NCEP reanalysis. Neither of the observational precipitation datasets examined here (Figs. A1c,d) can be invoked to support the strength of the precipitation dipole in the NCEP reanalysis, even though they both show a dipole anomaly in the North Atlantic. However, the heating anomaly in the European Centre for Medium-Range Weather Forecasts (ECMWF) Re-Analysis (ERA; Fig. A1a) is in good agreement with NCEP, and a very similar diagnosis can be obtained from the ERA data.

7. Summary and conclusions

The central finding of this paper is expressed in Fig. 5a, which shows that interactions between the zonal-mean flow anomalies and the climatological eddies make the dominant contribution to the maintenance of the NAO stationary waves (Fig. 3). This finding confirms the earlier demonstrations of Branstator (1984) and Ting et al. (1996) that zonal-eddy coupling is an effective mechanism for generating the stationary waves that accompany fluctuations of the Northern Hemisphere polar vortex. Using the full record of the NCEP reanalysis, we have extended their results by producing a complete and realistic diagnostic simulation of the stationary waves, so that the contribution of zonal-eddy coupling can be directly compared to the contributions of other forcing terms (heating and transients).

Our work differs from these studies in one important respect: we have tried to show that the zonal-eddy relationship is one of mutual adjustment, in which the zonal-mean flow also responds to the stationary wave anomalies. The zonal-mean flow anomalies of the NAO mode appear to interact with the climatological stationary waves to produce anomalous stationary waves, which further interact with the climatological eddies to generate momentum fluxes that help maintain the zonal-mean anomalies (Fig. 5d).

It should be emphasized that the positive feedback between zonal-mean and eddy flow anomalies is not so strong that they maintain each other exclusively, without any contribution from the other forcing terms. Heating and transients make some contribution to the stationary waves, and transients provide some momentum to the $\tau$ anomalies (DeWeaver and Nigam 2000, Fig. 9). In particular, the response to heating is important for the midtropospheric temperature perturbations over western Europe (Fig. 9). Ideally, the zonal-mean and eddy flow anomalies should be modeled together, so that their mutually adjusted response to the other forcing terms can be determined. This sort of modeling would be required, for example, to test Hurrell’s (1995b) claim that the NAO circulation anomalies are driven by vorticity transients from the Atlantic storm track. Unfortunately, our attempts to model the zonal-mean and eddy components together have not so far been entirely successful (Figs. 6 and 7).

The zonal-eddy coupling described here depends on the presence of strong climatological stationary waves and is thus specific to the north polar vortex in winter. Yet similar vortex fluctuations occur in the Southern Hemisphere and in simple models with zonally symmetric climatologies, as discussed in the introduction. In cases other than the Northern Hemisphere winter, the fluctuations are always found to be driven by transients (e.g., Karoly 1990; Limpasuvan and Hartmann 2000), with no significant input from anomalous stationary wave momentum fluxes. Given this substantial difference in the dynamics of the Northern and Southern Hemisphere annular modes, it is remarkable that they share a common meridional structure. Evidently, the factors that determine the central latitude and meridional scale of the annular $\tau$ perturbations are more general than the mechanisms discussed here. The search for these factors is clearly a priority for future research.

In light of these structural similarities, it is also natural to ask whether there are any substantial differences between annular fluctuations driven by transients and those described here that are associated with strong zonal mean flow—stationary wave interactions. In particular, would an AGCM with weak climatological stationary waves produce qualitatively different vortex fluctuations than an AGCM with strong stationary waves?

One possible difference involves the extent to which the vortex perturbations extend into the stratosphere. In Yu and Hartmann’s (1993) simple model, the stratospheric $\tau$ anomalies accompanying the Southern Hemisphere annular mode are much weaker than the anomalies of the NAO mode shown here (Fig. 7a). The observed Southern Hemisphere annular mode also has a weaker connection to the stratosphere throughout most of the year, although Thompson and Wallace (2000) have identified a strong connection during November. In our NAO composites, stationary wave fluxes provide momentum to the lower-stratospheric $\tau$ perturbations, a result that depends on the presence of robust clima-

Fig. 9 (Continued) 10 gpm, with dark (light) shading for positive (negative) values in excess of 10 gpm. In (e), contour interval is 0.1 K day$^{-1}$, with dark (light) shading for positive (negative) values in excess of 0.1 K day$^{-1}$ and zero contours suppressed. Heating in (e) differs from heating in Fig. 1a due to truncation and interpolation to R15L15 resolution, and because heating used to force the model is set to zero for $\sigma > 0.8$, as discussed in section 3.
to logical stationary waves. Of course, there is no guarantee that upper-level forcing will produce an upper-level response. A better understanding of how the vertical profile of $\pi$ depends on the vertical profile of the forcing is also a subject for future research.

Another potential role for zonal-eddy coupling involves the NAO’s response to extratropical SSTs. Rodwell et al. (1999) have reproduced the observed low-frequency variation of the traditional wintertime NAO index by forcing an AGCM with 50 yr of observed SSTs, a result that supports earlier claims of air–sea interactions in the North Atlantic (e.g., Palmer and Sun 1985; Grötzner et al. 1998). In our view, the dominant role of zonal-eddy coupling in the North Atlantic sector could have important implications for air–sea interactions in the region. For a strong air–sea interaction to occur, the atmospheric response forced by an SST anomaly should be positioned so that it can have further interactions with the SST anomaly. The diagnostic simulation presented here suggests that the position of the atmospheric response could be strongly influenced by zonal-eddy coupling, so that zonal-eddy coupling could have a controlling influence on the strength of air–sea interactions. Our results also imply that the NAO’s response to SST anomalies in an AGCM integration could be sensitive to the representation of the Atlantic sector stationary waves in the AGCM climatology.

In short, the zonal-eddy dynamics described here could be important for establishing causal relationships between the NAO and processes occurring in the ocean, the stratosphere, and other components of the climate system. Ultimately, it is these relationships that will determine the extent to which we can predict NAO or AO fluctuations and assess their role in climate change.

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APPENDIX A

Heating and Rainfall Comparisons

It could be argued that the model-generated heating fields in Fig. 2 are somewhat subjective, since they are determined by model parameterizations, and different heating estimates could be obtained simply by changing the parameterizations. The residual heating should be less model dependent, since the residual calculation uses analyzed winds and temperatures that are determined at least in part by the available observations. However, the close correspondence between residually diagnosed and model-generated heating could be an indication that the residual heating is also strongly influenced by the assimilating model. The model is used to provide first guess fields for the analysis, and these fields are especially influential in poorly sampled regions like the North Atlantic.

To test for model biases, we compared the residual heating from the NCEP reanalysis (Fig. 2a) with the residual heating calculated from the reanalysis of the ECMWF. The ECMWF reanalysis, commonly referred to as ERA (ECMWF Re-Analysis, described by Gibson et al. 1997), was available at NCAR for the years 1979–93 on a 6-hourly basis at $2.5^\circ \times 2.5^\circ$ resolution on 17 levels. The ERA winds and temperatures are determined at the analyzed winds and temperatures that are determined at the parameterizations. The residual heating should be less model dependent, since the residual calculation uses analyzed winds and temperatures that are determined at least in part by the available observations. However, the close correspondence between residually diagnosed and model-generated heating could be an indication that the residual heating is also strongly influenced by the assimilating model. The model is used to provide first guess fields for the analysis, and these fields are especially influential in poorly sampled regions like the North Atlantic.

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The NCEP precipitation anomalies are shown in Fig. A1b. Consistent with the movement of the Atlantic storm track, an area of reduced precipitation extends across the Atlantic from the Carolinas to the Iberian Peninsula, while enhanced precipitation occurs over the seas to the north between Greenland and the European coast. The area of reduced precipitation extends over land, crossing southern Europe to the Black Sea.

In Figs. A1c,d we compare the NCEP anomalies with precipitation anomalies taken from two sources: DaSilva et al.’s (1994) version of the Comprehensive Ocean Atmosphere Data Set (COADS), and the observations-only version of the Climate Prediction Center Merged Analysis of Precipitation (CMAP/O) by Xie and Arkin (1997). We used COADS data for the period 1958–89 and CMAP/O data for 1979–98. These datasets are generated without forecast models and are not subject to model biases. On the other hand, it is extremely difficult to measure precipitation over the oceans, and neither dataset can be considered definitive.

Like NCEP, COADS (Fig. A1c) shows a precipitation dipole in the North Atlantic with a maximum positive anomaly of 0.8 mm day$^{-1}$ southwest of Iceland, although precipitation anomalies are generally weaker and less extended in COADS than in NCEP. The southern cell of the dipole has an extreme value of about $-0.6$
a) ERA heating Jan. 1979 – Dec. 1993 0.1K day$^{-1}$

b) NCEP precip. Jan. 1958 – Feb. 1998 0.2mm day$^{-1}$


mm day$^{-1}$, compared to $-0.8$ mm day$^{-1}$ in NCEP, and does not reach the American coast. The northern cell has a much smaller northward extension into the Norwegian Sea than in NCEP, and the two datasets disagree strongly concerning conditions in the Davis Strait/Labrador Sea region, where COADS shows relatively large precipitation values that are not present in the NCEP data. The precipitation estimates in COADS are taken from ship reports of weather conditions that are converted into precipitation rates by a regression formula. Ship reports provide reasonable coverage in the shipping lanes that cover the southern cell of the dipole, but the seas to the north of Iceland are not well sampled. Data coverage has increased over time, but the more recent years contain a documented fair weather bias, as ships change course to avoid inclement weather.

The CMAP/O precipitation in Fig. A1d also shows a dipole in the North Atlantic, but the dipole is much weaker and less coherent here than in either COADS or NCEP. The best agreement between CMAP/O and NCEP is in the southern cell of the dipole, especially in its extension over land. The northern cell is weak and disconnected in CMAP/O, and there is a tendency for bullseye features along the coast. CMAP/O is generated entirely from rain gauge measurements and satellite data, of which the former are unavailable and the latter unreliable over the extratropical oceans. These deficiencies are acknowledged by Xie and Arkin, who emphasize the value of their dataset for tropical applications.

**Tropical heating and rainfall anomalies**

Our discussion of precipitation and heating anomalies has overlooked a variety of tropical and subtropical features, including rainfall anomalies over Brazil, the tropical North Atlantic, the Sahel, and the Indian and Pacific Oceans. These features are particularly interesting in light of previous work suggesting links between the NAO and the Tropics. For example, Namias (1972) found an association between blocking activity off the Newfoundland coast and rainfall in northeast Brazil. Also, the east--west dipole that covers the Indian Ocean and the Western Pacific in the CMAP/O rainfall in Fig. A1d is strongly suggestive of the Madden--Julian oscillation (MJO), and Ferranti et al. (1990) have shown evidence that the NAO can be excited by the MJO (see also Weickmann and Sardeshmukh 1994).

While these tropical connections are intriguing, they are not particularly strong. For the CMAP/O dataset, which shows the strongest tropical connections, the correlation with the NAO index is between 0.2 and 0.3 over most of the features of interest. It is entirely possible that the NAO has a small but consistent effect on the Tropics, or that the MJO and other tropical convection anomalies excite some monthly and interannual NAO variability, but more work is needed to establish the physical significance of the links to tropical rainfall in Fig. A1.

![Column-mean PBL diffusion](image)

**APPENDIX B**

**Boundary Layer Description**

Using the definitions below, the linearized vertical diffusion terms in the temperature ($T_s$), vorticity ($\zeta_s$), and divergence ($D_s$) equations can be written:

$$\frac{\partial}{\partial \sigma} \left( \nu \frac{\partial T_s}{\partial \sigma} \right) = -\frac{\sigma}{H_s} D_s \Gamma_{s,0} \left( \frac{\sigma}{H_s} \frac{\partial \zeta_s}{\partial \sigma} \right)$$

Temperature and velocity are relaxed to zero at the surface using the boundary conditions:

$$\nu \frac{\partial}{\partial \sigma} \left( \nu \sigma \frac{\partial T_s}{\partial \sigma} \right) = -\frac{\sigma}{H_s} D_s \Gamma_{s,0} \left( \frac{\sigma}{H_s} \frac{\partial \zeta_s}{\partial \sigma} \right)$$

These formulas are applied by setting $T_s(\sigma = 1) = T_s(\sigma = \sigma_s)$ and $\zeta_s(\sigma = 1) = \zeta_s(\sigma = \sigma_s)$, where $\sigma_s$ is the lowest $\sigma$ level. We use an insulated boundary condition at the model top, but the PBL viscosity ($\nu$) vanishes to machine precision well below the top, so model solutions are not sensitive to the top boundary condition. Here

$$\nu = 11 \times [1 + \tanh(10\pi(\sigma - 0.925))]$$

is the vertical diffusion coefficient. The profile is taken from Nigam (1994).

$$H_s = RT_s(\sigma) g^{-1},$$

the scale height. Here $R$ is the gas constant, $g$ gravitational acceleration, and $T_s(\sigma)$ the global mean of the climatological temperature for the level.

$$\Gamma_{s,0} = 9.77 \text{ K km}^{-1},$$

the dry adiabatic lapse rate.

$$\Gamma_{s,0} = 4.0 \text{ K km}^{-1},$$

the saturated adiabatic lapse rate, taken from Nigam (1994).

$$C_d = 3.0 \times 10^{-3},$$

the surface drag and sensible heat exchange coefficient.

$$|V| = 5.0 \text{ m s}^{-1},$$

the surface wind speed.
\[ \kappa = R C_p^{-1} \Gamma \Gamma_d^{-1}, \] where \( C_p \) is the specific heat at constant pressure.

The column heating due to PBL diffusion is plotted in Fig. B1 for the simulation in Fig. 3b. There is general agreement with the reanalysis PBL heating in Fig. 2d, with cooling over Scandinavia and some heating over Davis Strait and the Labrador Sea.

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