Do Stationary Waves Drive the Zonal-Mean Jet Anomalies of the Northern Winter?

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ABSTRACT

The role of zonal-mean zonal flow (u) perturbations in generating anomalous stationary waves has been acknowledged since the 1939 study by Rossby and his collaborators. However, the dynamical mechanisms, which in turn produce the u anomalies, are still not well understood. Here, the authors examine the forcing of u anomalies in the NCEP–NCAR 40-yr Reanalysis by regressing the zonal-mean zonal momentum budget against the leading empirical orthogonal function (EOF) of monthly December–January–February u in the domain covering 30°S to 90°N. The authors find that momentum fluxes arising from the interaction of climatological and anomalous stationary waves constitute the primary source of zonal-mean zonal momentum for the leading u EOF, which resembles the zonal index fluctuations discussed by Rossby, Namias, and others. When combined with previous studies that show the generation of stationary waves by u anomalies, the results presented here indicate a cooperative dynamical relationship between the u and stationary wave anomalies associated with the zonal index—a relationship in which each is both a source of and a response to the other.

The role of stationary waves in driving u anomalies is further examined for the canonical Northern Hemisphere teleconnection patterns. The authors use a rotated principal component analysis of 200-mb geopotential height to identify the u anomalies associated with the North Atlantic oscillation (NAO), the Pacific–North American (PNA) pattern, and an El Niño-related pattern. The NAO and PNA pattern are both accompanied by midlatitude u anomalies resembling EOF1. However, the two do not contribute equally to the leading u EOF: the NAO accounts for 64% of the variance, and the PNA pattern accounts for about 10%. Furthermore, the NAO clearly shows coherent expansion and contraction of the entire polar vortex, but the PNA pattern does not. The time series of the NAO may thus be a better indicator of the expansion and contraction of the polar vortex than are indices based on the zonally averaged circulation.

1. Introduction

Fluctuations in the midlatitude zonal-mean zonal wind (u) are implicated in much of the winter climate variability of the Northern Hemisphere. This fact was recognized by the pioneers of long-range weather forecasting, who were particularly interested in u fluctuations associated with the expansion and contraction of the polar vortex (e.g., Namias 1950; Rossby and Willett 1948; Wallace and Hsu 1985). More recently, Ting et al. (1996) claimed that such u fluctuations are more influential than El Niño in explaining the winter-to-winter variability of Northern Hemisphere surface air temperature (SAT). Furthermore, Thompson and Wallace (1998) find a strong resemblance between the changes in Northern Hemisphere SAT over the past 30 yr and the SAT pattern that accompanies short-term expansions and contractions of the north polar vortex (see also Hurrell 1995; Thompson et al. 2000). An understanding of the dynamics of u variations may thus prove essential in explaining the recent warming of the northern land-masses.

Northern Hemisphere u fluctuations are generally accompanied by large-amplitude stationary waves, and much of the climate change that occurs during these fluctuations is actually brought about by the stationary waves, through their anomalous meridional advection of temperature and moisture, and their effects on the storm tracks (e.g., Klein 1983). Recognizing the importance of the stationary waves, Rossby and his collaborators (1939) developed their well-known dispersion relationship, which showed how the waves could be thought of as a response to the u anomalies. The generation of stationary waves in response to u fluctuations has also been demonstrated by more recent diagnostic modeling studies, both for simulated (Kang 1990; see also Nigam and Lindzen 1989) and observed (Branstator 1984; Hoerling et al. 1995; Ting et al. 1996) circulations. In fact, Ting et al. suggest that u fluctuations are the most prominent source of wintertime stationary wave activity in the northern extratropics. All of this work, however, leaves open the question of how the u anomalies are generated in the first place.

To address this question, we calculated the zonal-
mean zonal momentum budget for the leading empirical orthogonal function (EOF) of monthly mean wintertime \( \pi \) variability in the domain covering 30°S to 90°N and calculated the corresponding budgets for the zonal indices of Ting et al. and Rossby et al. The budgets identify the roles of transients, anomalous stationary waves, and Eulerian-mean meridional motions in sustaining and limiting the \( \pi \) anomalies. In these budgets, momentum fluxes produced by the interaction between anomalous and climatological stationary waves are the most prominent momentum source for the \( \pi \) anomalies. Furthermore, Ting et al. have shown that these \( \pi \) anomalies produce a large stationary wave response in their diagnostic model. Our results thus suggest that the zonal-mean flow–stationary wave relationship demonstrated in the diagnostic modeling studies is best described as a *mutual* adjustment—an adjustment in which anomalous stationary waves are just as much a source of the \( \pi \) changes as a response to them.

The prominent role of stationary waves in maintaining \( \pi \) anomalies is somewhat unexpected, since strong \( \pi \) vacillations can occur in situations where climatological and anomalous stationary waves are weak or absent. The leading \( \pi \) EOF (EOF1) in our analysis (Fig. 1a) consists of a north–south dipole with centers at 35°N and 55°N—a pattern that Ting et al. identify with their zonal index and that Namias associated with Rossby's earlier index. Such north–south shifts of the zonal-mean jet, or zonal index fluctuations, have been simulated in a variety of models without orography or land–sea contrasts (e.g., Robinson 1996; Lee and Feldstein 1996; Yu and Hartmann 1993; James and James 1992). These studies obtain \( \pi \) fluctuations without any stationary wave anomalies and generally find that \( \pi \) vacillations are driven by momentum fluxes from baroclinic transients, a result anticipated by Namias. In some cases, transient eddy forcing produces surprisingly low-frequency vacillations, with periods ranging from 100–200
days in Yu and Hartmann to several decades in James and James.

Other studies (e.g., Thompson and Wallace 2000; Hartmann and Lo 1998; Nigam 1990; Karoly 1990; Kid-son 1985, 1986; Treberth 1984) have identified zonal index fluctuations in the Southern Hemisphere, where climatological and anomalous stationary waves are relatively weak. As in the above simple model studies, \( \pi \) fluctuations are driven by transients; this result holds both for the short-term (10-day \( e \)-folding) fluctuations discussed by Hartmann and Lo and for the interannual variability discussed by Karoly and Trenberth. Furthermore, none of these studies find qualitative differences between the zonal index fluctuations of the Northern and Southern Hemispheres, which suggests that zonal index anomalies are produced by the same mechanisms in both hemispheres. The observational studies of Thompson and Wallace (2000), Nigam (1990), and Kid-son (1985, 1986) found quite similar \( \pi \) fluctuations in the Northern and Southern Hemisphere midlatitudes.

On the other hand, the idea of mutual adjustment between \( \pi \) and stationary wave anomalies forms the core of Charney and DeVore’s (1979) simple model of atmospheric blocking. In this model, high and low zonal index states are maintained exclusively by cooperative feedback between \( \pi \) and the stationary waves, with no change in external forcing. Stationary waves contribute momentum to \( \pi \) through “form drag” as they interact with idealized orography, a process similar to the one described here, although in our case anomalous stationary waves interact with climatological stationary waves rather than with orography. Like orographic form drag, the anomalous–climatological stationary wave interaction is a linear function of the stationary wave anomaly. However, the prominent role of anomalous–climatological stationary wave interactions in maintaining \( \pi \) anomalies does not necessarily support Charney and DeVore’s contention that the \( \pi \) fluctuations result entirely from internal dynamics. Weickmann and Sardeshmukh (1994) and Weickmann et al. (1997) have shown how these stationary wave interactions can also serve as a mechanism for the driving of \( \pi \) anomalies by external forcing.

Weickmann and his collaborators discuss the driving of \( \pi \) anomalies by the interaction of climatological and anomalous stationary waves in the context of intraseasonal atmospheric angular momentum oscillations. They show that this interaction plays a prominent role in the generation and northward propagation of angular momentum perturbations by convection associated with the Madden–Julian oscillation (MJO). Of course, the propagating 30–70-day anomalies of their study may not have much influence on our monthly mean EOFs. However, the angular momentum perturbations examined by Weickmann and Sardeshmukh show a clear zonal index pattern similar to our EOF1, maintained in part by the interaction of the climatological jets with a stationary wave pattern similar to the Pacific–North American (PNA) pattern of Wallace and Gutzler (1981).

An important point discussed by Weickmann and Sar-deshmukh is that the presence of climatological stationary waves allows a zonally asymmetric pattern of external forcing to produce a zonal-mean response by generating stationary Rossby waves that then interact with the climatological stationary waves to produce a \( \pi \) anomaly. The long-term persistence of midlatitude \( \pi \) anomalies has prompted researchers to propose a variety of slowly varying zonally symmetric external causes, such as solar variations (Willett 1948), volcanic aerosols (Robock and Mao 1992), and anthropogenic greenhouse warming (Perlwitz and Graf 1995). Given the role of anomalous–climatological stationary wave interactions, it is also possible that long-lived \( \pi \) variations are produced by zonally asymmetric forcing effects, such as the eddy components of tropical or midlatitude SST anomalies.

Apart from any dynamical considerations, some researchers have contested the physical significance of \( \pi \) fluctuations in the northern midlatitudes by arguing that such fluctuations are not preferred modes of atmospheric variability, but simply averages of zonal wind perturbations associated with unrelated regional modes. For instance, while Ting et al. show that their zonal index is related to other regional modes, they also claim that the wind fluctuations in the two sectors are not well correlated with each other. In such a case, a zonal \( \pi \) anomaly would most likely result from the intensification of either the Atlantic or Pacific jet, each one driven by different regional mechanisms, rather than the action of a single hemispheric mechanism that simultaneously affects both jets. The lack of correlation between zonal wind fluctuations at different longitudes is also cited by Wallace and Hsu (1985) as a reason for the abandonment of zonal index–based forecasting schemes in the 1950s.

To what extent does the Northern Hemisphere zonal index represent a coherent mode of variability affecting the entire polar vortex rather than an amalgam of unrelated regional phenomena? We address this question by examining the \( \pi \) anomalies associated with the leading modes of a rotated principal component (RPC) analysis of the 200-mb height field. The analysis shows that there are indeed independent contributions to the first \( \pi \) EOF, centered on the two ocean basins, from the PNA pattern and the North Atlantic oscillation (NAO). However, the two contributions are by no means equal: the NAO clearly shows the expansion and contraction of the entire polar vortex that is generally attributed to the zonal index, while the PNA pattern has a much more wavelike appearance and a much smaller correlation with our EOF1.

Differences in the appearance of the NAO and the PNA pattern are also reflected in their zonal-mean zonal momentum budgets. The budget decomposition for the NAO is nearly identical to the budget for the first \( \pi \)
EOF, while the budget of the PNA pattern shows substantial differences, including a significant role for nonlinearities. In light of these differences, we believe that the time series of the NAO is a better indicator of the expansion and contraction of the polar vortex than indices based on the zonally averaged circulation, which combine the variability of the NAO and the PNA pattern.

The remainder of this paper is divided into five sections. Section 2 describes the data used in the analysis and displays the leading \( u \) EOF and the \( \bar{u} \) patterns covariant with the zonal index of Ting et al. and the traditional zonal index discussed by Rossby and Namias. Section 3 defines the budget terms, showing the relative contributions of the linear and nonlinear momentum fluxes. Section 4 shows the budget for the leading \( u \) EOF together with the budgets for the two zonal indices. Section 5 shows the leading modes of an RPC analysis of the 200-mb height field and discusses the contributions of these modes to the leading \( u \) EOF. Discussion and conclusions follow in section 6.

2. Data and EOF analysis

a. Data description

All data in this study come from the National Centers for Environmental Prediction–National Center for Atmospheric Research (NCEP–NCAR) 40-yr Reanalysis (Kalnay et al. 1996). Monthly mean pressure level data were obtained at 2.5° × 2.5° resolution from http://wesley.wwb.noaa.gov/ncep_data/index_sgi62.html for the period of January 1958–February 1998. The monthly means are averages of four 6-h analyses for each day. In addition to monthly mean fields such as winds and geopotential heights, the online dataset contains a variety of transient flux statistics, including the meridional flux of zonal momentum by submonthly transients, which is required for the calculation of the momentum budgets. The transient momentum flux is available on 12 levels, from 1000 to 100 mb, while the monthly mean winds are available on 17 levels, from 1000 to 10 mb. To limit the size of the dataset, the longitudinal resolution was reduced from 2.5° to 5° by removing grid points corresponding to longitudes that are not divisible by 5.

The extent to which the data reflect biases in the assimilating model rather than the dynamics of the real atmosphere is an important issue in all observational diagnostic studies. As a test of our results, we repeated the budget calculation for the zonal index of Ting et al. (shown in Fig. 4) using data for the period of 1979–93 from the reanalysis of the European Centre for Medium-Range Weather Forecasts (Gibson et al. 1997). Despite the differences in assimilation methods and the different period of record, the budget (not shown) is quite similar to the corresponding NCEP–NCAR budget. The similarity is perhaps to be expected, since in our domain the largest discrepancies between available atmospheric datasets involve the divergent circulation in the Tropics (e.g., Trenberth and Olson 1988), while the transients and stationary waves that maintain the \( \bar{u} \) anomalies discussed here are part of the more robust rotational circulation in the northern extratropics.

b. EOF analysis

We use a standard EOF analysis to identify the dominant patterns of \( \bar{u} \) variability. The analysis is based on covariance rather than correlation, and it includes monthly mean \( \bar{u} \) anomalies in \( 30^\circ S–90^\circ N \) on 12 levels: 1000, 925, 850, 700, 600, 500, 400, 300, 250, 200, 150, and 100 mb. Latitudes south of \( 30^\circ S \) were removed to eliminate the variability associated with the Southern Hemisphere zonal index, which accounts for almost exactly the same fraction of global \( \bar{u} \) variability as its Northern Hemisphere counterpart, so that the two cannot be separated by EOF analysis.

The first EOF (EOF1), which explains 30.6% of the variance, is shown in Fig. 1a. The spatial pattern has a dipole structure in which a slackening of the winds at \( 35^\circ N \) is accompanied by enhanced westerlies around \( 55^\circ N \). A positive anomaly (high index state) thus corresponds to a shift of the westerlies toward higher latitudes, or a contraction of the polar vortex. The northern cell of the pattern has a strong connection to the stratosphere, which is reminiscent of the high-latitude stratospheric anomalies found in Nigam’s (1990) leading \( \bar{u} \) mode. This connection has prompted some authors, like Perlwitz and Graf (1995) and Kodera and Yamazaki (1994), to argue that stratospheric driving has a strong influence on tropospheric climate. The accompanying time series (Fig. 1b) shows variability on all timescales, with persistent anomalies in the winters of 1971/72 to 1975/76 and 1988/89 to 1992/93. Also, there is a tendency for negative values in the first decade of the record and positive values in the first half of the 1990s, consistent with the 30-yr strengthening of the polar vortex (e.g., Thompson et al. 2000; Thompson and Wallace 1998; Hurrell 1995).

Figure 1c shows the \( \bar{u} \) pattern covariant with the zonal index of Ting et al. (1996), which is formed by taking the difference in 500-mb \( \bar{u} \) between \( 35^\circ \) and \( 55^\circ \)N. The index (Fig. 1d) is shown here multiplied by \(-1\) for easier comparison with EOF1, with which it has a correlation of \(-0.79\). This strong correlation is to be expected, since the latitudes used to construct the index are the central latitudes of the EOF1 dipole. However, the comparison is still of interest because the Ting et al. index measures the \( \bar{u} \) pattern linked with the leading mode of stationary wave variability forced by zonal-eddy coupling in their diagnostic model, while EOF1 measures the strongest pattern of \( \bar{u} \) variability. The fact that the two patterns are so similar is an indication of the close dynamical relationship between the \( \bar{u} \) and stationary wave anomalies.
Figure 1e shows the patterns covariant with the traditional zonal index of Rossby et al. (1939), formed from the difference in zonal-mean sea level pressure (SLP) between 35° and 55°N. The index was intended to measure the mean strength of the westerlies in the intervening latitudes, which is proportional to the SLP difference under the geostrophic approximation and the assumption that the SLP fluctuations are equivalent barotropic. Despite the intentions of its authors, the Rossby et al. index is associated not with a simple midlatitude wind intensification, but with a dipole similar to that of the Ting et al. index. The correspondence between the patterns for the two indices is not exact, however, because the dipole for the Rossby index is found slightly to the south of the dipole for the Ting index, and a small additional negative cell appears near 75°N for the Rossby index. Also, for the Rossby index, the stratospheric connection in the northern cell of the dipole is much weaker than its counterparts for the Ting index. The correspondence between the indices is positive but relatively weak, as discussed below. In addition to the tendency and error terms discussed above, the variable \( R_i \) now contains a nonlinear term, \( N L_{R_i} \), given by

\[
NL_{R_i} = \sum \alpha_i \left( v_{an} \bar{\zeta}_a - \omega_a \frac{\partial u_{an}}{\partial p} \right). \tag{4}
\]

The expression \( NL_{R_i} \) is not shown separately in the momentum budgets because it was found to make a negligible contribution in comparison to the linear momentum flux terms in the large parentheses in (3). Of course, the dominance of the linear momentum fluxes over the nonlinear fluxes cannot be guaranteed a priori, and in section 5c we provide an example in which the two are equally strong (Figs. 10c.d). But given that \( NL_{R_i} \) is small, (3) can be regarded as a linear balance between four contributors to the momentum budget: the interaction between climatological and anomalous zonal-mean motions (the first set of terms in parentheses), including anomalous advection of planetary vorticity; the interaction between climatological and anomalous stationary waves (the second set of terms in parentheses); transient momentum fluxes; and the budget residual, which is dominated by boundary layer dissipation. While the term stationary wave is used here to refer to both the rotational and divergent components of the zonally asymmetric flow, the stationary wave contribution is, as one would expect, dominated by the rotational flow components.

Ideally, the zonal-mean zonal momentum budgets should exclude below-ground data and include a mountain pressure contribution from elevation differences between points where the pressure surfaces intersect the ground. However, in the budgets shown below, the dominant balance in the free atmosphere is between the Coriolis force and the momentum flux by transients and stationary waves, with the momentum fluxes occurring at altitudes that are largely above the highest surface.
elevations. The bogus underground data and the neglected mountain pressure term will thus have their impact mainly on the residual term, which has all of its amplitude at low levels. To determine the effects of the underground data and neglected mountain forcing, we recalculated the vertically integrated momentum budget residual for EOF1 in $\sigma$ coordinates. The $\sigma$-coordinate budget, which automatically excludes below-ground data, agrees with the pressure-level budget to within a few percent over most of the domain.

4. Budgets for $\bar{u}$ EOF1 and zonal indices

a. Momentum budget for $\bar{u}$ EOF1

Figure 2 shows the momentum budget for EOF1. It is clear from the figure that while transients (Fig. 2a) and stationary waves (Fig. 2b) both provide momentum to maintain the $\bar{u}$ anomalies (contoured in Fig. 2d), the stationary waves make a larger contribution, giving a maximum acceleration of $0.98$ m s$^{-1}$ day$^{-1}$, compared with $0.71$ m s$^{-1}$ day$^{-1}$ for the transients. Also, the stationary wave momentum forcing extends to higher levels than the transient forcing. This fact is significant because the positive $\bar{u}$ anomalies at high latitudes extend well into the lower stratosphere, whereas the vertical extent of synoptic transients and their momentum fluxes is limited by the tropopause. In the Southern Hemisphere, where stationary wave forcing is much weaker, the leading $\bar{u}$ EOFs have more limited connections to the stratosphere except in the month of November—the month when the southern polar vortex breaks down (Thompson and Wallace 2000).

The forcing by mean-meridional motions, shown in Fig. 2c, is almost entirely due to the Coriolis term $\bar{u}_x f$. In the northern positive cell of the $\bar{u}$ dipole (contoured in Fig. 2d), there is a well-defined thermally indirect circulation in which the Coriolis force acts on the mechanically generated jet perturbations to produce an Eulerian-mean meridional overturning motion in which momentum is transported down to the boundary layer where it can be effectively dissipated. The climatological $\bar{u}$ (not shown) has a node at $60^\circ$N, which marks the boundary between the Ferrel cell and the thermally direct polar cell. Thus, the positive phase of the index is accompanied by an expansion of the indirect Ferrel cell at the expense of the direct polar cell.

The budget residual (Fig. 2d) is strongly suggestive of Ekman layer dissipation, since it has large values in the boundary layer below the $\bar{u}$ anomalies and acts to oppose them. The southern cell of the $\bar{u}$ dipole also appears to be maintained by mechanical forcing due to transients and stationary waves, with the two playing more or less equal roles, balanced by the Coriolis term and Ekman layer dissipation. The budget terms in the southern cell are considerably weaker than those in the northern cell, as is the secondary circulation.

b. Column-mean forcing and $\bar{u}$ response

A notorious problem with budget studies such as this one is the lack of a clear dynamical relationship between the $\bar{u}$ anomalies and the forcing terms in the budget. The momentum deposited in the zonal mean by the eddies at upper levels is transported to lower levels in the Eulerian-mean secondary circulation, so the dynamical connection between the forcing and $\bar{u}$ response at upper levels is obscure. In the column mean, one might expect the eddy momentum forcing to produce a $\bar{u}$ response, but even then it may be difficult to determine
what $\tau$ response should be produced by a given forcing distribution. If surface drag balances the forcing by eddies, it seems quite likely that the boundary layer $\tau$ anomalies will have the same sign as the forcing. The bulk surface drag $\tau$ is usually obtained from $-\rho C_d |\bar{V}| u$, where $\rho$ is the surface air density, $C_d$ is the drag coefficient, and $|\bar{V}|$ is the surface wind speed. Thus, the anomalous zonal-mean surface drag depends on

$$
(\rho C_d |\bar{V}| u)_a = (\rho C_d |\bar{V}|) \bar{u}_a + (\rho C_d |\bar{V}|)^H u_a^H + \sum a_i (\rho C_d |\bar{V}|)_i u_a^c,
$$

so that a balance between momentum contributed by eddies and surface drag still does not lead to an exact relationship between forcing and $\tau$ response at the surface or in the column mean. The situation is further complicated if the eddy forcing is balanced by the mountain pressure term, which has no explicit dependence on the zonal wind at all. This term can be written as $p_w \partial h / \partial x$, where $p_w$ is the surface pressure anomaly and $h$ is the mountain height.

Despite these concerns, Fig. 3 shows that there is a strong empirical relationship between the column-mean forcing and $\tau$ response for EOF1. The curve marked with open circles is a pressure-weighted average of the terms in Figs. 2a–c, with the Coriolis force set to zero. The curve marked with filled circles is a pressure-weighted average of $\tau$, from 1000 to 10 mb, divided by 6 to give it the same scale as the forcing curve. The approximate agreement between the two curves suggests that in the column-mean, the $\tau$ anomalies are maintained by the convergence of momentum flux against a column-mean momentum dissipation with a timescale of 6 days, consistent with the classical Ekman spin-down time (e.g., Holton 1992).

It could be argued that it is the surface $\tau$ rather than the column-mean $\tau$ that is influential in balancing the eddy forcing through surface drag, so that the resemblance between the forcing and the column-mean response is an artifact of the barotropic vertical structure of the $\tau$ anomalies. The dashed line in Fig. 3 shows the zonal mean of the zonal wind at the surface, which is formed by interpolating the pressure-level zonal wind to the surface pressure using linear interpolation on the logarithm of pressure. The surface $\tau$, which is scaled by a factor of $2.5^{-1}$, agrees quite well with the column-mean $\tau$ except in the region of the stratospheric $\tau$ maximum between 60° and 70°N (Fig. 1a). The implied 2.5-day Rayleigh damping time is consistent with the boundary layer dissipation estimate provided by Valdes and Hoskins (1988). The surface $\tau$ does not, however, provide a better fit to the forcing curve. In northern latitudes, the discrepancy between forcing and $\tau$ response may be due to the influence of the mountain term. In lower latitudes, the difference may result from the other terms in the surface drag formula discussed above or the meridional variation of $(\rho C_d |\bar{V}|)_i$. Although a full investigation of the mechanisms that balance the eddy forcing terms and lead to the observed $\tau$ response is clearly warranted, it is beyond the scope of this paper.

c. Momentum budgets for zonal indices

Figures 4a–d show the budget for Ting et al.’s index, which is strongly similar to the EOF1 budget. The prominent role of stationary waves in driving the $\tau$ fluctuations associated with this index is especially significant because the index is designed to track the forcing of stationary waves by $\tau$ anomalies. Thus, Fig. 4 implies that the zonal-mean flow–stationary wave relationship
discussed by Ting et al. is a feedback in which $\bar{u}$ and stationary wave anomalies act to maintain each other.

Figures 4e–h show the momentum budget for the index discussed by Rossby and Namias. Again, stationary waves make a larger momentum contribution than transients. The differences between this budget and the EOF1 budget are consistent with the differences in the $\bar{u}$ patterns: there is a slight southward shift of the budget terms and less forcing at high elevations, consistent with the weaker stratospheric connection. Also, the Eulerian-mean meridional flow forcing and boundary layer dissipation have a third positive cell associated with the negative $\bar{u}$ anomalies in the high latitudes. The fact that stationary waves make the dominant contribution to the Rossby index as well as to our EOF1 suggests that stationary wave forcing is an important source of Northern Hemisphere $\bar{u}$ variability in general rather than being a process specific to EOF1. In fact, the momentum budgets of the first three EOFs of our $\bar{u}$ analysis (not shown), which together account for 65% of the variance in their domain, all show a substantial momentum contribution from stationary waves.
5. Asymmetric circulations associated with $\pi$ fluctuations

a. Regional contributions to $\pi$ forcing by stationary waves

What pattern of stationary waves drives the $\pi$ dipole of EOF1? To understand how the stationary waves affect $\pi$, we must consider both their shape and their placement relative to the climatological stationary waves with which they interact. Figure 5a displays the zonally asymmetric (eddy) components of the EOF1-covariant streamfunction at 250 mb, the level at which stationary wave forcing is largest, superimposed on the streamlines of the climatological December–February (DJF) flow. The corresponding linear momentum flux, $F_L = u^* v^* + u^* v^*$, is plotted in Fig. 5b. While covariant streamfunction anomalies occur throughout the Northern Hemisphere, the features in the Atlantic and European sectors are largely responsible for the momentum fluxes that drive the $\pi$ anomalies. For instance, over the North Atlantic, the eddy streamfunction indicates a westerly zonal wind anomaly in a...
Fig. 6. Rotated loading vectors (RLVs) for the first three modes of an RPC analysis of the 200-mb geopotential height field [(a), (c), and (e)], with covariant $\pi$ perturbations [(b), (d), and (f)]. Zero contours are suppressed in all panels and dark (light) shading represents positive (negative) values. The RLVs are displayed with a contour interval of 10 geopotential meters and covariant $\pi$ patterns are contoured at 0.5 m s$^{-1}$ intervals. The percentages given in the upper-right-hand corners of (a), (c), and (e) refer to the fraction of global 200-mb-height variance explained by each mode.

region where the climatological flow has a southerly component, so that $u^* v^* > 0$. Over western Europe, the eddy zonal wind anomaly is easterly while the climatological flow is somewhat northerly, so that again $u^* v^* > 0$. As discussed in section 6 (Fig. 11), the anomalous stationary waves can be thought of as changes in the tilt of the troughs and ridges that bring about the transport of momentum by the time-mean flow.

Some caution is in order when examining the local values of momentum fluxes such as those in Fig. 5b. In the zonal mean, a small northward or southward bias in the flux over a large area could outweigh the con-
FIG. 7. (a) Sea level pressure field covariant with the leading RPC mode in Fig. 6a. Contour interval is 1 mb, with dark (light) shading for positive (negative) values and zero contours suppressed. (b) Time series for the mode in Fig. 6a.

tribution of a few strong but localized centers. Thus, the figure may be overemphasizing the importance of the Atlantic and western European features discussed above. To assess the contribution of these features, we compared the zonal mean of the momentum flux in Fig. 5b with the zonal mean produced when all fluxes outside the region from 100°W to 60°E were set to zero. In the second case, the zonal mean is weaker by 20%–30%. The comparison shows that although other fluxes cannot be neglected, the bulk of the momentum flux is indeed occurring in the Atlantic and western European sectors.

b. EOF1 and the canonical modes of variability

Does EOF1 represent a preferred mode of atmospheric variability, or is it a synthetic pattern that reflects the influence of several localized modes? This question is examined in Fig. 6, which shows the leading modes of variability taken from an RPC analysis of the 200-mb geopotential height (Figs. 6a,c,e), together with the corresponding $\tilde{u}$ patterns (Figs. b,d,f). The RPC analysis is performed on the covariance matrix of the height field, which is Fourier interpolated onto an equal-area grid before the covariances are calculated. The first 10 eigenvectors of the covariance matrix are rotated using the varimax method (e.g., Barnston and Livezey 1987), although similar results can be obtained by rotating 8, 12, or 15 modes.

The leading mode of the RPC analysis clearly shows the kind of in-phase expansion and contraction of the entire polar vortex suggested by EOF1. The pattern has a negative cell over the polar cap surrounded by a belt of positive values with centers in the Atlantic and Pacific sectors. The covariant $\tilde{u}$ anomalies have almost exactly the same pattern as EOF1 (Fig. 1a), and the temporal correlation between this mode and EOF1 is 0.80. To the extent that EOF1 represents a preferred mode of the longitudinally dependent circulation, it clearly represents this mode.

The largest amplitudes of the mode occur in the Atlantic sector, with an opposition between Greenland and western Europe that is reminiscent of the NAO as well as the Arctic oscillation (AO) of Thompson and Wallace (1998, 2000). The association between this mode and the NAO and AO is confirmed in Fig. 7a, which shows the covariant sea level pressure (SLP). The covariant pattern closely resembles the leading EOF of SLP used by Thompson and Wallace to identify the AO, as well as the NAO-covariant SLP pattern of Hurrell (1995, Fig. 3). Although the SLP pattern in Fig. 7a has slightly more zonal symmetry than Hurrell’s pattern, it lacks the larger Pacific sector amplitude that distinguishes the AO from the NAO (cf. Thompson and Wallace 2000, their Fig. 1d, with Hurrell 1995, his Fig. 3b). The time series of the leading 200-mb height mode is shown in Fig. 7b to facilitate comparison between this mode and other indices. The correlation between this time series and the time series of the leading SLP EOF is 0.85, and its correlation with Hurrell’s NAO index, obtained from http://www.cgd.ucar.edu/cas/climind/nao.monthly.html, is 0.93.

The second mode (Fig. 7c) has a correlation of 0.80 with the Niño-3.4 SST index and shows the anticyclones that straddle the equator in the eastern Pacific during El Niño winters. The $\tilde{u}$ pattern for this mode consists primarily of a subtropical jet enhancement of the sort that one would expect from ENSO events (Nigam 1990; Hoerling et al. 1995).

The third mode has all of the characteristics of the PNA pattern (Wallace and Gutzler 1981, Fig. 16), with a three-celled wave train crossing North America from the Aleutian Gulf to the mid-Atlantic coast of the United States. Although this teleconnection pattern does not represent a coherent fluctuation of the polar vortex, the corresponding $\tilde{u}$ pattern has a dipole structure and a barotropic vertical profile similar to EOF1. The correlation between EOF1 and the PNA mode is $-0.34$. Sim-
ilar correlations hold for the zonal index of Ting et al., which correlates with the NAO mode at ~0.64 and with the PNA mode at 0.38. The relative importance of PNA variability for their index thus is somewhat greater than for our EOF1.

Examination of the NAO- and PNA-covariant $\overline{u}$ patterns thus shows that while EOF1 is strongly associated with the NAO, it actually represents a combination of NAO and PNA variability. In our view, the inclusion of PNA variability in any index designed to measure the simultaneous in-phase fluctuation of the entire polar vortex is undesirable, given the wavelike structure of...
the PNA pattern. It would thus seem that the time series of the NAO mode is a better zonal index for the Northern Hemisphere than any index based on the zonally averaged circulation.

c. Zonal–eddy dynamics of the NAO and PNA modes

The spatial patterns of the NAO and PNA modes suggest that there may be differences in their zonal–eddy dynamics, and we begin our consideration of these differences by examining the stationary waves associated with the two modes. The 250-mb eddy streamfunction anomalies for the modes, as well as the corresponding anomalies for EOF1, are plotted in Fig. 8. Figure 8a shows the EOF1 anomalies (same as in Fig. 5a, but centered on the date line), and Fig. 8b shows the stationary waves for the linear combination of the NAO and PNA modes that provides the best fit to Fig. 8a. The NAO- and PNA-covariant eddy streamfunctions are shown independently in Figs. 8c and 8d. Together, the NAO and PNA modes account for about 75% of the variance of EOF1, and a comparison of Figs. 8a and 8b reveals that all of the EOF1 eddy streamfunction anomalies in the northern extratropics are captured in the NAO–PNA composite. Further comparison of Figs. 8c and 8d shows that, although the PNA mode explains only 10% of the global EOF1 variance, it accounts for most of the amplitude of the EOF1 stationary waves over the central Pacific and western North America, while the NAO mode accounts for the EOF1 features over the Atlantic and Eurasian sectors.

From Fig. 5 we know that the stationary wave momentum fluxes that sustain the EOF1 \( u \) anomalies occur primarily in the Atlantic and European sectors. Since the NAO mode accounts for nearly all of the EOF1 stationary wave activity in these sectors, we expect that stationary wave forcing will play the same role in the zonal-mean zonal momentum budget for the NAO mode as it does for EOF1. This expectation is confirmed in Fig. 9, which displays the zonal-mean zonal momentum contributed by transients (Fig. 9a) and linear stationary wave fluxes (Fig. 9b) for the NAO. The high degree of similarity between these terms and the corresponding terms for EOF1 (Figs. 2a and 2b) also holds for the mean meridional term and the budget residual (not shown).

The spatial pattern of the PNA mode suggests a different zonal–eddy relationship. Unlike their NAO counterparts, the PNA stationary waves are not well positioned to generate northward momentum fluxes by interacting with the climatological stationary waves. The streamfunction dipole in the central Pacific (Fig. 8d) implies a strong zonal wind anomaly in the exit region of the East Asian jet, which extends and retracts the jet but has a relatively small impact on the poleward flux of zonal momentum in the region. An examination of the linear momentum fluxes for the PNA mode over the American sector (not shown) shows that there, too, the linear fluxes are weaker and less organized than the linear fluxes associated with the NAO mode in the Atlantic and European sectors. On the other hand, for a \( u \) anomaly of a given magnitude, the accompanying stationary waves would be much stronger for the PNA mode than for EOF1 or the NAO mode (Fig. 6). The relative importance of the nonlinear momentum flux \( u^*v^* \) should thus be greater for the PNA mode than for the NAO mode.

Figure 10 compares the relative strengths of the linear and nonlinear stationary wave momentum fluxes associated with the PNA and NAO modes. The linear and nonlinear fluxes are given by \( F_L = \pi^*u^* + \pi^*v^* \) and \( F_{NL} = a_i u_i a_i v_i \), respectively. Whereas the linear flux dominates the nonlinear flux by a factor of 8 for the NAO mode (Figs. 10a and 10b), the two contributions are nearly equal for the PNA mode (Figs. 10c and 10d). The figure thus provides further evidence of the strong qualitative differences between the zonal–eddy dynamics of the two modes, differences that support our view that PNA variability should be excluded from diagnoses of the zonal–eddy dynamics of polar vortex fluctuations.

6. Concluding remarks

Simply stated, the stationary waves associated with the monthly mean Northern Hemisphere zonal index and
the related NAO mode maintain the corresponding $\overline{u}$ anomalies by modulating the tilt of the Atlantic jet. Figure 11 shows the 250-mb streamlines of the Atlantic sector flows that would occur for extreme positive (Fig. 11a) and negative (Fig. 11b) anomalies of the NAO mode. The streamlines were constructed by adding and subtracting twice the 250-mb wind anomalies covariant with the NAO mode (Fig. 7a) to and from the climatological winds. In the positive phase, the Atlantic sector ridge has a pronounced southwest-to-northeast tilt, so that the northward flux of zonal momentum ($u^*v^*$) is larger than its climatological value. In the negative phase, the ridge line has a somewhat less pronounced tilt in the opposite direction. Although the anomalies in the figure are extreme, a negative anomaly of this size did occur in February 1960; January 1984, February 1989, and January 1993 had equally large positive anomalies.

It must be emphasized that the momentum budget does not establish a causal relationship between the $\overline{u}$ and stationary wave anomalies. Only if the stationary waves are generated independently of the $\overline{u}$ anomalies, as in Weickmann and Sardeshmukh's (1994) MJO diagnosis, can it be said that the stationary waves cause the $\overline{u}$ anomalies. On the other hand, the stationary waves could be produced by the $\overline{u}$ anomalies themselves, through their effect on the refraction of stationary waves produced by the climatological vorticity sources (e.g., Nigam and Lindzen 1989). The relationship between the zonal-mean and eddy flow anomalies would then be one of mutual adjustment, and no causality could be ascribed to either flow component in the time mean. Both scenarios are consistent with the momentum budgets shown here, but the studies of Branstator (1984) and Ting et al. (1996) provide strong support for mutual adjustment, because they show that the $\overline{u}$ anomalies help to maintain the stationary waves.

Our discussion so far suggests that the circulation anomalies associated with the NAO and the zonal index are manifestations of internal atmospheric dynamics that would occur even in the absence of external forcing. However, external forcing may still be implicated in the...
low-frequency behavior of the anomalies. Although the simple atmospheric models of James and James (1992) and Yu and Hartmann (1993) produced low-frequency oscillations based solely on internal dynamics, such low-frequency behavior has not been reported in realistic atmospheric general circulation models (AGCMs) forced by climatological boundary conditions. The two most obvious aspects of the low-frequency index behavior of Figs. 1b and 7b are the long-term upward trend, which has been associated with the warming of the northern landmasses, and the persistence of anomalies of the same sign from winter to winter.

Perlwitz and Graf (1995) find a similar long-term trend in a leading mode of the combined variability of the stratosphere and troposphere, which resembles our NAO mode. They ascribe the trend to anthropogenically induced global warming, arguing for a “top-down” influence in which greenhouse warming first affects the strength of the stratospheric polar vortex, which then modifies the tropospheric circulation and surface temperatures. The mechanism for top-down control is taken from work by Kodera and collaborators (e.g., Kodera 1994; Kodera et al. 1996), who find a zonal-eddy feedback mechanism similar to the one discussed here in their studies of stratosphere—troposphere interactions. These authors claim that the zonal-eddy feedback mechanism makes the troposphere sensitive to changes in the stratospheric polar vortex, because changes in the stratospheric polar vortex can modulate the propagation of stationary waves, which in turn create tropospheric $\overline{u}$ anomalies.

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tex have not traditionally been identified as a response to CO$_2$ doubling in global warming simulations, but positive AO trends were reported in the studies of Shindell et al. (1999) and Fyfe et al. (1999). Shindell et al. obtained their result only when enhanced resolution was used in the stratosphere, but Fyfe et al. used very limited stratospheric resolution. In both cases, the trends are weak in comparison with observations.

Even if the stratospheric effects of global warming are behind the long-term trend, global warming cannot easily explain the multiyear persistent periods that occur for both phases of the NAO mode. Rather, the persistence suggests that the mode is affected by conditions at the lower boundary, such as snow cover (Kodera and Koide 1997), sea ice, or SSTs. The recent study of Rodwell et al. (1999) presented a simulation in which most of the low-frequency behavior of Hurrell’s NAO index is reproduced by an AGCM forced with observed SSTs.

Despite the success of this simulation, the influence of SSTs is still a matter of some debate. Several studies (e.g., Deser and Timlin 1997; Battisti et al. 1995) have shown that atmospheric flow anomalies in the North Atlantic sector have a strong effect on the underlying ocean so that the ocean can be called upon to "remember" the NAO mode. These oceanic memories can only explain persistence if they can be recalled by the atmosphere at a later time, however, and attempts to show an atmospheric response to midlatitude SST anomalies have had mixed results. For example, Palmer and Sun (1985), Lau and Nath (1990), and Latif and Barnett (1996) found a robust atmospheric response, but Lau and Nath (1994), Kushnir and Held (1996), and Bladé (1997) got very limited responses.

As a first step in understanding the processes that lead to the low-frequency behavior of the NAO mode, we have undertaken a dynamical diagnosis of the NAO-related stationary waves using a steady linear primitive equation model. The diagnosis, to be reported in a forthcoming paper (DeWeaver and Nigam 2000), provides a detailed description of the forcing of the stationary waves associated with the NAO mode. As in previous studies, zonal-mean flow changes provide the dominant forcing to the stationary waves, although transients and diabatic heating are also important. More research is needed, however, to determine how these forcing terms are influenced by conditions in the stratosphere and at the lower boundary.

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REFERENCES


