Mechanisms of Zonal Index Variability in an Aquaplanet GCM

Steven Feldstein

Earth System Science Center, The Pennsylvania State University, University Park, Pennsylvania

Sukyoung Lee

Department of Meteorology, The Pennsylvania State University, University Park, Pennsylvania

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ABSTRACT

Data from a 2025-day Geophysical Fluid Dynamics Laboratory aquaplanet GCM integration are used to examine the temporal evolution of the zonal index, defined as the principal component of the first EOF of the zonally and vertically integrated absolute angular momentum. This EOF represents meridional displacements of the subtropical jet. Positive (negative) values of the zonal index correspond to poleward (equatorward) displacements of the subtropical jet and are referred to as the high (low) zonal index.

Composites of various quantities are used to examine the temporal evolution of both the zonal-mean zonal winds and the eddy fields associated with either index. The high index is initiated by an enhancement in the equatorward wave activity propagation, which causes the subtropical jet to move in a poleward direction. The low index is led by a weakening in the equatorward propagation, which results in an equatorward displacement of the subtropical jet. More importantly, for the high index, the corresponding increase in eddy forcing is confined to a brief period near onset, resulting in rapid growth and slow decay of the zonal wind anomaly. This suggests that the high index state is “impulsively” forced by the eddies rather than maintained by eddy–zonal mean feedback. In contrast, for the low index, a reduction in the eddy forcing extends throughout the entire persistent episode, although the weakest eddy forcing occurs during onset. The authors believe that such forcing causes the low index anomaly to persist for a longer period of time than the high index anomaly and also results in a similar growth and decay rate for the low index. Furthermore, case studies show that the onset for both indices is associated with wave breaking. For the high (low) index, this wave breaking takes the form of filamentation (blocking).

1. Introduction

The zonally averaged circulation in midlatitudes has been known for many years to exhibit low-frequency variability. A measure of the strength of the midlatitude westerlies between 35°N and 55°N, known as the zonal index, was introduced by Rossby (1939). Namias (1950) showed that the midlatitude circulation fluctuated irregularly between high and low zonal index states, the high (low) index being characterized by a poleward (equatorward) shift in the latitude of maximum zonal-mean westerly winds. He named these zonal wind fluctuations as the index cycle. Further support for these properties came from Lorenz (1951) who found strong negative correlations between the sea level pressure at 35°N and 55°N. Interest in the index cycle then declined during the 1950s as its usefulness for weather prediction became increasingly questionable (Wallace and Hsu 1985). This was because the index cycle implied simultaneous growth and decay of eddies at widely spaced longitudes, that is, a uniform wave train across the entire hemisphere, which does not typically occur in either hemisphere. Instead, as indicated by recent studies such as Lee and Held (1993), synoptic-scale eddies are most often organized into coherent baroclinic wave packets.

In recent years, there has been renewed interest in low-frequency variability of the zonally averaged flow, (e.g., Rogers and van Loon 1982; Kidson 1985, 1986, 1988; Nigam 1990; Karoly 1990; Robinson 1991; Lyons and Hundermark 1992; Yu and Hartmann 1993). These observational and modeling studies, most of which employ empirical orthogonal functions (EOFs), find that the dominant type of low-frequency fluctuation involves latitudinal displacements of the westerly jet. These fluctuations in the zonally averaged flow reveal many of the characteristics of Namias’s zonal index cycle, but they do not show recurring sequences of events suggested by the term “cycle.”

There are several reasons for the renewed interest in zonal index behavior. Perhaps, the most important reason is because of the hope that a better understanding
of low-frequency variability of the zonally averaged flow, which is the simplest form of low-frequency variability, may give insight into the properties of the more complex zonally asymmetric low-frequency anomalies (Wallace and Gutzler 1981). Although not an entirely separate issue, another reason is because of the role the zonally averaged flow plays in determining the structure of the planetary-scale stationary waves (e.g., Branstator 1984; Kang and Lau 1986; Nigam and Lindzen 1989).

Although the temporal and spatial characteristics of zonal index variability are now well described, much of the fundamental dynamics of this phenomenon remain poorly understood. Both the observational study of Karoly (1990) and the modeling studies of Robinson (1991) and Yu and Hartmann (1993) suggest that the transient eddy momentum flux convergence maintains the anomalous zonal mean flow when it is displaced toward either side of its time-mean latitude. Also, Yu and Hartmann (1993) found the momentum flux convergence to be important for the transition from one index to the other. Using idealized life cycle experiments, they argue that nonlinear baroclinic life cycles (Simmons and Hoskins 1978) may play a role in the maintenance of the zonal index anomalies.

However, basic questions remain about the dynamical mechanisms associated with zonal index variability: 1) What is the structure of the eddies throughout the zonal index evolution? In particular, is eddy–zonally mean feedback a robust feature? 2) Are there any differences in the zonal index characteristics between a subtropical jet and a polar front or eddy-driven jet? 3) What determines the timescale for the onset and persistence of the zonal index? In this study, we address these and other questions about zonal index variability with data from an aquaplanet general circulation model (GCM). This model has the sea surface temperatures (SSTs) specified to be independent of longitude and symmetric across the equator. The advantage of such a model is that it allows one to examine the zonal index in a model with simple boundary conditions and yet retain all the physical parameterizations of a full GCM. Furthermore, as the boundary forcing in the model is independent of time, it allows one to isolate the zonal index variability arising from the model’s internal dynamics and to exclude the effects of time-varying boundary conditions, which are expected to play a role in the atmosphere.

The model characteristics and data analysis techniques are presented in section 2. This is followed by a brief description of the model’s climatology in section 3. The results are described in section 4. A discussion is presented in section 5 and the conclusions are given in section 6.

2. Data analysis and procedure

The data examined in this study were generated from a 2025-day run of an aquaplanet Geophysical Fluid Dynamics Laboratory (GFDL) GCM. This model has a rhomboidal-30 horizontal resolution, 9 sigma levels, and a lower boundary that is a flat ocean surface. A fixed annual mean insolation is used, that is, there is no seasonal cycle. This model includes a full radiation package, predicted clouds, and a moist convective adjustment scheme. The identical GCM has been used in studies by Lee and Held (1993) to examine the characteristics of baroclinic wave packets and by Feldstein and Lee (1995) to study the intraseasonal evolution of atmospheric angular momentum. As the model generates all data on sigma surfaces and the results to be presented will be on pressure surfaces, logarithmic interpolation from sigma to pressure coordinates is performed.

There are many possible definitions that could be used for the zonal index. In this study, we define the zonal index as the principal component of the first EOF of the hemispheric absolute angular momentum $M(\phi, t)$, where $\phi$ is latitude and $t$ time. The absolute angular momentum, $M(\phi, t)$, is equal to the vertical and zonal integral of $p_i 4\Omega \cos \phi + u a \cos \phi$, where $p_i$ is the surface pressure, $\Omega$ the earth’s angular velocity, $a$ the earth’s radius, and $u$ the zonal wind speed. As we will see, the structure of this EOF will indeed correspond to meridional displacements of the subtropical jet. Positive (negative) values of the zonal index are defined to correspond with poleward (equatorward) latitudinal shifts of the subtropical jet. We will refer to positive (negative) values of the zonal index as high (low) index states. Also, as will be discussed later, the results of this study are insensitive to other definitions of the zonal index.

In order to define persistence formally, we follow the procedure adopted by Horel (1985) and Mo (1986). The pattern correlation $r$ at day $t$ and lag $\tau$ is defined as the spatial linear correlation

$$r(t, \tau) = \frac{\langle m'(\phi, t)m'(\phi, t+\tau) \rangle}{\sigma[m'(\phi, t)]\sigma[m'(\phi, t+\tau)]},$$

where $m(\phi, t)$ is the deviation of $M(\phi, t)$ from its corresponding time mean at that latitude, $m' = m(\phi, t) - \langle m(\phi, t) \rangle$, $\sigma^2[m'(\phi, t)] = \langle (m'(\phi, t))^2 \rangle$, and the angle brackets denote a meridional average over one hemisphere.

The following objective criterion, based on $r(t, \tau)$, is used to define a persistent episode: an episode is defined as being persistent if for 5 or more consecutive days the pattern correlations $r(t, \tau)$ and $r(t + 1, \tau)$ are greater than or equal to a particular threshold value. In other words, a persistent anomaly is defined to take place if $r(t, \tau)$ and $r(t + 1, \tau) \geq r_c$, for $\tau = 1$ to 5. The threshold value $r_c$ is chosen as the 95% significance level. In order to determine $r_c$, it is necessary to know the number of degrees of freedom. This can be obtained from Fisher’s Z-transformation:
\[ Z(t, \tau) = \frac{1}{2} \log \left( \frac{1 + r(t, \tau)}{1 - r(t, \tau)} \right). \] (2)

If \( Z \) is normally distributed, then the number of degrees of freedom, \( N \), can be determined by calculating the variance of \( Z \), which is equal to \((N - 3)^{-1}\).

The definitions for onset and decay are based on the above definition for persistence. The onset day will be defined as the first day of the persistent episode. The decay day is defined in a slightly different way. We redo the above patterns correlations, but progress in the opposite direction, that is, calculate \( r(t, \tau) \) and \( r(t - 1, \tau) \) for \( \tau = -1 \) to \(-5\). The decay day corresponds to the last day of this persistent episode. In almost every case, we find that the same episode is selected from both the forward and backward pattern correlations, and the onset day precedes the decay day. In addition, a middle day is defined as the day corresponding to the midpoint between the onset and decay days.

Persistent episodes for the high (low) index phase are then determined by the additional requirement that at day \( t \) the zonal index is greater (less) than \( 1.0 \pm 1.0 \) standard deviation, and that the magnitude of the principal components of the remaining EOFs are less than that for EOF1. This combination of using both statistically significant pattern correlations and large values for the zonal index ensures that the absolute angular momentum field is always dominated by the first EOF. In some cases, it is found that the zonal index crosses this one standard deviation threshold before the onset day. For these cases, the onset day is shifted backward to the day at which the zonal index first exceeds one standard deviation. This shifting procedure is performed to make sure that at the onset day the zonal index has a similar value for each case. The results to be presented in this study are essentially the same whether or not this shifting procedure is performed. Also, when an onset day occurs within 15 days of the previous decay day, those onset days are discarded and are regarded as occurring within the persistent episode beginning on the previous onset day. An analogous procedure is also applied for discarding decay days that take place within 15 days of the following onset day.

In this study, unless stated otherwise, composites of various quantities are performed relative to the onset day. The quantities to be shown in the composite analyses include Eliassen–Palm (EP) cross sections and various fields such as zonal and meridional wind. The statistical significance of the composites presented in this study were estimated with a two-sided \( t \)-test. For each of these composites, the \( t \) values showed similar spatial patterns that exceeded the 99% confidence level. These composites will be complemented with case examples, which are particularly illuminating as they show smaller-scale structures related to wave breaking that are filtered out by the compositing procedure. We also note that all data used in this study will remain unfiltered.

3. Climatology of the GCM

Before examining the characteristics of the GCM’s zonal index, it will be beneficial to first examine the climatology of several quantities and to compare these quantities to those observed in the atmosphere. The model’s climatological zonally averaged zonal and meridional winds, averaged over both hemispheres, are shown in Figs. 1a and 1b, respectively. Compared with the average of the March and September observations (Oort and Rasmussen 1971), which approximates mean equinox conditions, the model’s 205-mb jet is approximately in the same location, but is about 10 m s\(^{-1}\) stronger than that observed. The time-mean me-
ridional winds, in agreement with observations, indicate three meridional cells. As can be seen, the time-mean Hadley cell extends to about 25° (the poleward end of the Hadley cell is regarded as the latitude where the mean meridional wind vanishes), about 5° equatorward of the subtropical jet maxima.

Due to the proximity of the jet to the Hadley cell, one would expect that this jet takes on some of the characteristics of a subtropical jet (Krishnamurti 1961). Thus, in this model, the evolution of the zonal index should also be influenced by subtropical jet dynamics and its associated Hadley cell. As we will see, this is indeed the case in this GCM. In contrast, the idealized model studies of Robinson (1991) and Yu and Hartmann (1993) place their jet farther poleward, and thus their jet is expected to exhibit properties closer to that of an eddy-driven, polar front jet. However, since there is a single climatological jet both in this GCM and in the idealized models of Robinson (1991) and Yu and Hartmann (1993), these jets are expected to exhibit mixed characteristics of subtropical and eddy-driven jets to varying degrees. In this study, the role of the subtropical jet for the zonal index will be emphasized.

4. Results

a. Description of the zonal index

The temporal variation of the absolute angular momentum field is examined with an EOF analysis. As discussed in section 2, we define the zonal index as the principal component of the first EOF of the hemispheric absolute angular momentum field. This EOF analysis is performed separately for both hemispheres. The absolute angular momentum, which is both vertically and zonally integrated, is separated into 80 latitudinal bands, each band of equal meridional width of 2.25°.

The first two Northern Hemisphere EOFs are shown in Fig. 2. The first EOF, whose spatial structure corresponds to that of the zonal index, accounts for 60% of the total variance, whereas the second EOF comprises only 16% of the total variance. It can be seen that EOF1 has a dipole structure with its node located at 32°, which is close to the latitude of the model’s subtropical jet maximum. Because the variance of the absolute angular momentum is found to be dominated by its \( p, \mu \cos \phi \) contribution, for a sufficiently large zonal index, EOF1 corresponds to meridional displacements of the subtropical jet. On the other hand, the second EOF shows that its maximum value is located very close to the latitude of the subtropical jet maxima. This indicates that EOF2 represents a strengthening and weakening of the subtropical jet. Most other zonal index studies that use EOFs also find these characteristics for EOF2.

As expected, because of the model’s cross-equatorial symmetry, the corresponding Southern Hemisphere EOF analysis (not shown) indicates that the first two EOFs have a meridional structure that is almost identical to those for the Northern Hemisphere. The corresponding variance for these Southern Hemisphere EOFs is 65% (16%) for EOF1 (EOF2). Therefore, to approximately double the sample size, the composite calculations to be presented in this study combine data from both the Northern and Southern Hemispheres. Following the procedure for determining persistence, as described in section 2, it is found that there are 28 (29) persistent episodes for the high (low) index.

We also examined the sensitivity of this procedure to the variable used in the EOF analysis. This was accomplished by redoing the EOF analysis with relative angular momentum (the vertical and zonal integral of \( p, \mu \cos \phi \)), vertically and zonally averaged zonal wind, and 205-mb zonally averaged zonal wind. These EOF analyses, together with their pattern correlation analyses, select most of the same persistent episodes as was found with absolute angular momentum.

The length of time for persistence of either phase of the zonal index can be obtained from composites of the zonal index (see Fig. 3) relative to the onset day, which is defined as lag 0. If we define the length of the persistent episode to be the time period for which the amplitude of the zonal index exceeds one standard deviation, then the high and low index events persist for 7 and 10 days, respectively. Furthermore, it can be seen that the low index anomaly decays more slowly than the high index anomaly.

Using the above definitions for onset and persistence, we next look at composites of the anomalous vertically and zonally averaged zonal winds at several lags both prior to and after onset, for both the high and
Fig. 3. Lag composites of the zonal index relative to the onset day (lag 0) for the high index (positive values) and low index (negative values). The straight dashed lines indicate the one standard deviation level.

low zonal index (Figs. 4a and 4b). As expected, there are anomalies that correspond to the two extrema in EOF1. In a manner consistent with the evolution of the zonal index (see Fig. 3), both high and low index anomalies exhibit rapid growth from 2 days before onset until 2 days after onset, followed by a slow decay of the anomaly. However, it is important to note that this compositing procedure selectively favors those days for which the zonal wind tendency is large. As a result, for flows that lack a preferred period, the composite zonal wind tendency is enhanced near the onset day, and smoothed out away from the onset day. Thus, in order to check whether the picture presented in Figs. 4a and 4b of rapid growth followed by slow decay is an artifact of the compositing procedure, we examine the composite anomalous vertically and zonally averaged winds based on both the middle and decay days. For the low index, we indeed find that the above temporal asymmetry is an artifact of the compositing procedure, as there is symmetric growth and decay on opposite sides of the middle day (Fig. 4d), and slow growth followed by rapid decay on either side of the decay day (Fig. 4f). On the other hand, for the high index, the growth is still more rapid than the decay in the middle day composite (Fig. 4c), and symmetric growth and decay is found in the decay day composite (Fig. 4e). As the middle day composite best characterizes the temporal symmetry of either index, the high index does indeed undergo rapid growth and slow decay, whereas for the low index, growth and decay occur at about the same rate. Furthermore, the middle day composites find, in a manner consistent with Fig. 3, that the low index persistence is longer than that for the high index. The significance of these differences in the growth and decay of either index will be discussed in the next subsection.

The remaining composites to be presented in this study will be based on the onset day. We choose the onset day, rather than the middle or decay day, in order to emphasize the flow evolution at the time when it is changing most rapidly. However, as discussed above, it should be borne in mind that composites based on the onset day exaggerate the rate of growth relative to that of decay.

Throughout the evolution of the zonal index, the vertical structure of the zonally averaged zonal wind anomalies is found to remain equivalent barotropic with largest amplitude in the upper troposphere. Examples of this vertical structure are shown in Figs. 5a and 5b for two days after onset (recall that the composite zonal index attains its largest amplitude at two days after onset). Composites of the total zonally averaged zonal wind field are shown for 2 days after onset in Figs. 5c and 5d. As expected, the zonal index does indeed correspond to latitudinal shifts of the subtropical jet, with the difference in the latitude of the subtropical jet maxima between the two indices being about 4° at 205 mb. At this level, the maximum zonal winds are also stronger for the low index composite.

The power spectrum for the zonal index indicates a red noise structure (not shown). A red noise power spectrum was also found by Feldstein (1995) for the zonal index in another GFDL GCM that included continents, topography and seasonally varying sea surface temperatures.

b. Eddy forcing

As it is anticipated that much of the zonal index evolution is associated with eddy-zonal mean flow interaction, we examine EP cross sections at various days relative to onset. Additional information on EP flux diagnostics can be found in Edmon et al. (1980). The EP flux vector, in quasigeostrophic form for pressure coordinates, can be written as $\mathbf{F} = (0, F^{(\phi)}, F^{(p)})$, where

$$F^{(\phi)} = -a \cos \phi \left( \frac{\bar{v}}{\bar{u}} \frac{\partial \bar{\psi}}{\partial \phi} \right)$$

$$F^{(p)} = a \cos \phi \left( f_0 \frac{\partial \bar{\psi}}{\partial p} \right),$$

where $v$ is the meridional wind, $\theta$ the potential temperature, $f_0$ a constant Coriolis parameter, $a$ the earth’s radius, and $\theta_0$ the horizontal mean potential temperature. The asterisk denotes a deviation from a zonal average, and the overbar indicates a zonal average. In the strictest sense, $\mathbf{F}$ is parallel to the group velocity vector only for small amplitude disturbances. The corresponding quasigeostrophic form for the zonally averaged zonal wind tendency equation is

$$\partial \bar{u} / \partial t = \nabla \cdot \mathbf{F} + f_0 \bar{\psi},$$
where $\overline{v^y}$, the transformed meridional wind, satisfies

$$\overline{v^y} = \overline{v} - \partial / \partial p \left( \frac{v \partial \theta}{\partial \theta / \partial p} \right).$$

(6)

Eliassen–Palm cross sections are shown for the high and low index at onset in Figs. 6a and 7a, respectively.

Both figures are dominated by upward wave activity propagation in midlatitudes, with stronger vertical propagation occurring in the high index case. In the upper troposphere, on the poleward side of the subtropical jet, the wave activity propagation is weak for both cases. On the other hand, on the equatorward side of the subtropical jet, there is a substantial amount of

Fig. 4. Composites of the anomalous zonally and vertically averaged zonal wind relative to the (a) high index onset day, (b) low index onset day, (c) high index middle day, (d) low index middle day, (e) high index decay day, and (f) low index decay day. Lag zero corresponds to the onset day in (a) and (b), the middle day in (c) and (d), and the decay day in (e) and (f). Contour interval is 0.4 m s⁻¹. Solid contours are positive, dashed contours negative, and the zero contour is omitted.
equatorward wave activity propagation for the high index, whereas for the low index, the wave activity propagation is again weak. The corresponding EP flux divergence field shows that the spatial pattern of the eddy forcing is rather similar for both zonal indices, since the eddy forcing for either index accelerates (decelerates) the zonal wind on the poleward (equatorward) side of the subtropical jet. The primary distinction between the two zonal indices is that the high (low) index eddy forcing is stronger (weaker) relative to that for the time-mean eddy forcing.

The anomalous EP flux vectors and their divergence are shown in Figs. 6b–d at various lags relative to onset for the high index. At 5 days before onset, anomalous upward EP flux vectors indicate that the poleward heat flux is enhanced throughout much of the midlatitudes. The largest amplitude anomalous vectors occur at about 830 mb. At 2 days before onset, the largest amplitude anomalous vectors have moved upward and evidence of enhanced equatorward propagation can be seen. By onset, most of the anomalous wave activity propagation is equatorward and concentrated in the upper troposphere. Also, at this time, the largest anomalous EP flux divergence takes place, which acts to accelerate (decelerate) the subtropical jet on its poleward (equatorward) side.

For the low index (see Figs. 7b–d), at 5 days before onset, downward anomalous EP flux vectors indicate a weakening of the poleward heat flux in midlatitudes. Also, as for the high index, the largest anomalies occur in the lower troposphere. At 2 days before onset, the largest anomalous vectors have moved upward, and
poleward anomalous wave activity propagation is found in the upper troposphere. By the onset day, most of the anomalous wave activity propagation is poleward and confined to the upper troposphere. At this time, as for the high index, the anomalous EP flux divergence is greatest and acts to decelerate (accelerate) the subtropical jet on its poleward (equatorward) side. Thus, the anomalous vectors for the low index show similar characteristics, but of opposite sign, to those for the high index.

In order to obtain a clearer picture of the role of the eddy forcing, we examine the anomalous lower tropospheric (from 990 to 660 mb) meridional heat flux (Figs. 8a and 8b) and the anomalous vertically integrated meridional momentum flux convergence (Figs. 8c and 8d). For the high index, the anomalous heat flux changes sign immediately after onset and shortly thereafter shows no organization. The anomalous momentum flux convergence, which must cause the subtropical jet to move poleward, is largest at onset and becomes much smaller in magnitude and disorganized only two days after onset, which is the time at which the zonal wind anomaly attains its maximum amplitude. The corresponding middle day composite of the anomalous momentum flux convergence (not shown) also finds the eddy forcing to be essentially confined only to those days prior to the maximum anomaly. Recalling that the anomalous zonal winds decay slowly over a period of about 20 days (see Fig. 4a or 4c), a qualitative picture for the
high index emerges, which suggests that the eddy forcing can be viewed as "impulsive," and that the anomaly is not maintained by the eddies during the persistence and decay stages.

On the other hand, the composite eddy fluxes for the low index reveal a different picture. As for the high index, both the anomalous heat flux and the anomalous momentum flux convergence attain their largest values prior to the day of the maximum zonal wind anomaly, but these anomalous fluxes prolong for the remainder of the persistent episode (similar properties are found for the composite fluxes based on the middle day, except that the fluxes show greater symmetry about either side of the middle day). It is this eddy forcing that most likely accounts for the similar growth and decay rates for the low index anomaly, and the fact that the low index persists for a greater period of time than the high index. Recalling from Fig. 7a that the low index eddy forcing is weaker than the time-mean eddy forcing, one can characterize the low index as a lengthy period of weak forcing. However, it is unclear whether this extended weaker eddy forcing can be explained by an eddy-zonal mean feedback.

The anomalous zonal mean storm track, that is, the anomalous high pass (2–10 day) 515-mb eddy geopotential height variance, is primarily positive (negative) during the onset to the high (low) index, reaching its extreme value on the poleward side of the time-mean storm track. For the high index, this results in a slight poleward movement of the storm track maximum together with a 20% increase in its strength. On the other hand, for the low index, the storm track maximum remains at its time-mean latitude with a 20% reduction in strength.

With the stronger eddy heat and momentum fluxes during the onset of the high index, as compared to the
Fig. 8. Lag composites of the anomalous heat flux for (a) the high index and for (b) the low index, and of the anomalous momentum flux convergence for (c) the high index and (d) the low index. Contour interval in (a) and (b) is $0.15 \text{ K m s}^{-2}$, and in (c) and (d) is $1 \times 10^{-9} \text{ m s}^{-2}$. Solid contours are positive, dashed contours are negative, and the zero contour is omitted.

low index, one expects that the Hadley and Ferrel cells are stronger for the high index, which is indeed found to be the case (not shown). Furthermore, in a manner consistent with the strength of the Hadley circulation, the high index possesses more precipitation in equatorial regions than the low index. A comparison of the

Fig. 9. Composite anomalous $\gamma n^2$ and Eliassen–Palm vectors at the onset day for (a) the high index and (b) the low index. Contour interval is $6.0 \times 10^{-9} \text{ m s}^{-2}$. Shading is above $1.2 \times 10^{-9} \text{ m s}^{-2}$ in (a) and below $-1.2 \times 10^{-9} \text{ m s}^{-2}$ in (b). The maximum vector lengths are illustrated in the figure. Solid contours are positive, dashed contours are negative, and the zero contour is omitted.
anomalous $^1\gamma u^T$ with the anomalous $\nabla \cdot F$ (cf. Figs. 9a and 9b with Figs. 6d and 7d) indicates that at onset for either index, in both the subtropics and midlatitudes, the anomalous $\nabla \cdot F(\gamma u^T)$ dominates in the upper (lower) troposphere. Thus, for example, during the onset to the high index, the poleward shift of the upper troposphere subtropical jet is caused by the anomalous $\nabla \cdot F$ pair, whereas the poleward shift of the jet in the lower troposphere is accomplished by the anomalous $\gamma u^T$ pair. Identical arguments, but of opposite sign, apply for the low index.

c. Case examples

This subsection presents the flow morphology of two representative cases, one for the onset to the high index, and the other for the onset to the low index. These cases are illustrated with maps of potential temperature $\theta$ on the 2-PVU potential vorticity surface and horizontal wind vectors on the same surface. The technique used for generating these potential temperature plots is discussed in detail in Lee and Feldstein (1996a). In the present study, we use these potential temperature maps to identify particular atmospheric features that coincide with the onset of either index.

The potential temperature maps for the high index case are shown in Fig. 10, together with the zonal mean zonal wind. This is illustrated for days 1046 through 1050, where day 1048 is the onset day. On day 1046, there are two particular troughs that we will examine closely. One trough is located along the subtropical jet at 80° latitude and 20° longitude, and the other trough is associated with a midlatitude disturbance located at 145° longitude and 45° latitude. During the next two days, the trough along the subtropical jet propagates about 40° eastward and the midlatitude trough propagates both southward about 10° latitude and also slightly eastward. As a result, on day 1048, the northernmost of the two troughs is located to the east of the southern trough. The resulting horizontal phase relation between these two troughs is oriented in such a manner as to transport momentum poleward. This poleward momentum transport is consistent with the poleward movement of the zonal mean subtropical jet. By day 1050, the two troughs have separated. Also, in the vicinity of the southernmost trough, the poleward movement of the subtropical jet has increased the anticyclonic shearing deformation. The enhanced strain field stretches the trough into a broad filament, which then breaks.

The example for the low index case is shown in Fig. 11. This example covers the period from day 1151 to 1155, where onset corresponds to day 1153. On day 1151, two days before onset, we can see that compared to day 1046, which is two days before onset for the high index (see Fig. 10a), the amplitude of both the midlatitude eddies and those eddies along the subtropical jet are much weaker. Following the discussion in subsection 4b, it was noted that weaker midlatitude eddy fluxes result in a deceleration (acceleration) of the zonal mean zonal winds on the poleward (equatorward) side of the subtropical jet. Due to the equatorward movement of the subtropical jet, cyclonic shear in this region is enhanced, increasing the northwest–southeast tilt of the ridge at 130° longitude and 50° latitude (see day 1153). This ridge continues to amplify and cuts off by day 1155, forming a “block.” One also finds a large block at 360° longitude at day 1153, but the role of this particular block in the low index evolution is unclear. These blocks continue to persist through to day 1160 (not shown) and propagate westward, which is characteristic for blocks in the atmosphere.

5. Discussion

As is well known, when eddies of midlatitude origin propagate into the Tropics and then break, the zonally averaged flow is accelerated (decelerated) at the latitude of the wave source (sink) through eddy momentum flux convergence (divergence). For the high index, as is shown in the example of Fig. 10a, this breaking occurs between 15° and 30° latitude. According to idealized modeling studies (Held and Phillips 1987; Feldstein and Held 1989), this zonal-mean flow deceleration, or wave drag, is located in the vicinity of the wave’s critical latitude, that is, where the wave’s phase speed matches that of the background zonally averaged flow.

Thus, based on the previously shown EP flux diagrams and case examples, we can view the high (low) index as coinciding with enhanced (reduced) wave drag on the equatorward side of the time-mean jet maximum and increased (decreased) eddy momentum flux convergence on the poleward side of the time-mean jet maximum. Note that for the low index, the positive anomaly in the eddy momentum flux convergence near 25° latitude (see Fig. 8d) is not due to increased eddy momentum flux convergence, but is due to decreased eddy momentum flux divergence.

As discussed above, the subtropical jet plays an important role in the dynamics of the zonal index in this paper. This contrasts the results of Lee and Feldstein (1996b), who examined the zonal index characteristics of a purely eddy-driven jet in a two-layer quasigeostrophic $\beta$-plane model. In that model, Lee and Feldstein (1996b) find that the onset to the high and low index are antisymmetrical, as the high (low) index corresponds to equatorward (poleward) wave activity propagation across the eddy-driven jet. Such wave activity propagation characteristics typically does not oc-

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$^1$ We use $\gamma u^T$ instead of $f\omega^T$, where $\gamma = [f - (1/a \cos \phi)(\partial a / \partial \phi)]$, because the quasigeostrophic form becomes less accurate in low latitudes.
Fig. 10. Case example for the high index illustrating the potential temperature and horizontal wind vectors on the 2-PVU potential vorticity surface. Days 1046 through 1050 are shown, where day 1048 is the onset day. The day is indicated in the top left corner of each frame. The corresponding panels to the right of the potential temperature maps show the zonally averaged zonal wind on the same day (solid line) together with the climatological zonally averaged zonal wind (dashed line). The maximum vector lengths are illustrated in the figure. Contour interval is 5 K and shading is below the 310-K contour. All contours are positive.

cur within a subtropical jet, as that jet is bounded on its equatorward side by a Hadley cell, where wave activity is much diminished. As a result, most eddies that interact with the subtropical jet originate in midlatitudes and propagate in an equatorward direction across the jet. It is the fluctuation in the strength of this equatorward wave activity propagation that causes the zonal index behavior associated with the subtropical jet.

The temporal evolution of the baroclinicity parameter, \(0.31 f (\partial \bar{u} / \partial z) N^{-1}\) (not shown), suggests that the anomalous meridional heat fluxes associated with either sign of the zonal index, are not driven by the anomalous baroclinicity. This is because there is a contemporaneous relationship between the anomalous baroclinicity parameter and the anomalous vertically and zonally integrated zonal wind (see Fig. 4), while the
anomalous meridional heat flux maximum occurs several days earlier. Therefore, at least in this GCM, it is unlikely that normal mode baroclinic instability is responsible for the anomalous eddy forcing.

In another study that used data from the same GCM run, Lee and Feldstein (1996a) found that eastward propagating wave packets were a prevalent feature of the model circulation. Following the procedure in Lee and Feldstein (1996a), wave packets were identified and then phase-shifted so that their maximum amplitude occurred at a common longitude. This technique was applied to each of the persistent events to obtain a composite at onset of various statistics such as wave packet amplitude, heat, and momentum fluxes. For both indices, it is found that the composite heat and momentum fluxes are confined within local wave packets. As expected from the EP flux diagnostics described earlier, the composite heat and momentum fluxes are greater for the high index than for the low index. The main point is that even in this idealized GCM, the eddy forcing of the zonal index is localized rather than zonally uniform at a given time, and the lack of zonally uniform eddy growth and/or decay does not deter the usefulness of the zonal index concept.
6. Conclusions

This study examined fundamental characteristics of the high and low zonal index in an aquaplanet GCM. The general picture that emerges for the onset to either index is as follows. During the onset to the high (low) index, there is an enhancement (weakening) of the vertical component to the wave activity flux that is followed by an increase (decrease) in the equatorward wave activity flux in the upper troposphere. For the high index, this enhancement of the eddy forcing results in a poleward movement of the subtropical jet, and for the low index, this weakening of the eddy forcing causes the subtropical jet to move toward the equator. These results show that the zonal index behavior associated with the subtropical jet is driven by variation in the strength of the equatorward propagation of midlatitude eddies. This is the case, even though the time-mean subtropical jet is maintained to lowest order by the conservation of absolute angular momentum in the upper branch of the Hadley cell and dissipated by wave drag. Such behavior contrasts with that for the zonal index associated with an eddy-driven jet (Lee and Feldstein 1996b), where both the time-mean jet and the zonal index are driven by the eddies.

For the high index, the enhancement in the eddy forcing is found to be confined to a very short period near onset, resulting in the zonal wind anomaly growing rapidly and decaying slowly, with the anomaly not being maintained by the eddy forcing. This suggests that one can view the high index as being "impulsively" forced. On the other hand, for the low index, although the weakest eddy forcing occurs near onset, the state of reduced eddy forcing remains throughout the entire persistent episode. We believe that it is for this reason that the low index zonal wind anomaly persists for a greater period of time than the high index anomaly, and grows and decays at a similar rate. However, it is unclear whether this prolonged low index persistence is due to eddy–zonal-mean feedback.

An open question is what determines the timescale for the decay of either index anomaly. For the high index, in the absence of eddy forcing during its decay, it is likely that the timescale is determined by dissipative processes, such as boundary layer dissipation (daily surface stresses were not saved in this GCM run, so the role of surface friction could not be accurately determined). To lowest order, one expects a three-way balance in the vertically integrated zonal momentum equation between 1) the zonal wind tendency, 2) the eddy momentum flux convergence, and 3) surface friction. To the extent that the surface drag can be approximated as a large residual of a balance between the first two terms, a calculation of the anomalous residual term finds that the high index anomaly does indeed decay through surface friction. Similarly, for the low index, equating the anomalous residual term with surface friction shows that its decay is also strongly influenced by surface friction; however, the decay timescale is enhanced by the persistent weak eddy forcing.

The structure of the eddies during the onset to the low index was examined with potential temperature maps. Typically, it was found that the high (low) zonal index is associated with filamentation (blocking). It was also found that during the onset to most high index events, there was an equatorward propagating midlatitude disturbance whose trough was located to the northeast of a trough on the subtropical jet. This "trough phasing," or wave merger (Lai and Bosart 1988; Gage and Bosart 1990; Hakim et al. 1995) also appeared to play an important role in the enhancement of the poleward momentum flux. As for the high index of this model, the onset to either index in the quasi-geostrophic model of Lee and Feldstein (1996b) frequently involves two waves that interact with one another. However, in their model, unlike the aquaplanet GCM, for both the high and low index, there were many cases for which two troughs and/or two ridges merged. This trough phasing, and often merger, is an interesting phenomenon, with potentially important implications for low-frequency variability of the zonal mean flow. It certainly warrants additional theoretical study.

Since the onset to the high and low zonal indices are associated with variation in meridional wave activity propagation, one may expect some relationship with the two paradigms of baroclinic life cycle behavior found by Thornroft et al. (1993). The onset to the high index indeed resembles their "LC1" life cycle, exhibiting a cutoff low on the equatorward side of the jet. However, the onset to the low index, characterized by a cutoff high on the poleward side of the jet, differs from their 'LC2' life cycle, which possesses a long-lived cutoff low on the poleward side of the jet. However, a recent Southern Hemisphere observational study found LC1 (LC2) behavior associated with the high (low) zonal index (Hartmann 1995). This suggests that although LC1 characteristics are typical for the high index, the kind of wave breaking that coincides with the low index can be more variable and probably depends upon the detailed characteristics of the background flow.

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