Kinematics of Eddy–Mean Flow Interaction in an Idealized Atmospheric Model

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ABSTRACT

The authors analyze atmospheric variability simulated in a two-layer baroclinic $\beta$-channel quasigeostrophic model by combining Eulerian and feature-tracking analysis approaches. The leading mode of the model's low-frequency variability (LFV) is associated with the irregular shifts of the zonal-mean jet to the north and south of its climatological position accompanied by simultaneous intensification of the jet, while the deviations from the zonal-mean fields are dominated by propagating anomalies with wavenumbers 3–5. The model's variability is shown to stem from the life cycles of cyclones and anticyclones. In particular, synthetic streamfunction fields constructed by launching idealized composite-mean eddies along the actual full-model-simulated cyclone/anticyclone tracks reproduce nearly perfectly not only the dominant propagating waves, but also the jet-shifting LFV. The composite eddy tracks conditioned on the phase of the jet-shifting variability migrate north or south along with the zonal-mean jet. The synoptic-eddy life cycles in the states with poleward (equatorward) zonal-jet shift exhibit longer-than-climatological lifetimes; this is caused, arguably, by a barotropic feedback associated with preferred anticyclonic (cyclonic) wave breaking in these respective states. Lagged correlation and cross-spectrum analyses of zonal-mean jet position time series and the time series representing mean latitudinal location of the eddies at a given time demonstrate that jet latitude leads the storm-track latitude at low frequencies. This indicates that the LFV associated with the jet-shifting mode here is more dynamically involved than being a mere consequence of the random variations in the distribution of the synoptic systems.

1. Introduction

Earth’s atmosphere exhibits turbulence characterized by a broad range of spatial and time scales. Day-to-day weather variations in midlatitudes are dominated by traveling low- and high-pressure systems—cyclones and anticyclones—which have radii up to a few thousand kilometers. In the Northern Hemisphere, these synoptic eddies form two major storm-track regions in the North Atlantic and North Pacific (Gerber and Vallis 2007; Swanson 2007; Williams et al. 2007). The storm tracks, while composed of individual synoptic eddies, vary themselves in their extent and location on time scales exceeding a few days, thereby contributing to what we will hereafter call the atmospheric low-frequency variability (LFV). Dominant modes of LFV can be identified via space/time bandpass filtering. The application of filters hinders dynamical interpretation of LFV, in a sense that an LFV pattern so obtained, while representing the average over an ensemble of true solutions of the governing equations, may not itself be such a solution. The same applies to the climatological midlatitude jet, which is obtained by taking the long-term time mean of a highly variable atmospheric flow. In this paper, we combine two complementary methodologies—the one that relies on identifying individual cyclones/anticyclones and their tracks and the one based on space/time filtering—to obtain a consistent kinematic picture of midlatitude LFV in an idealized atmospheric model.

The filtering approach provides an appealing and efficient way of analyzing complex atmospheric variability,
with the storm tracks defined as the regions of enhanced variation of the 2.5–6-day bandpass-filtered height fields (e.g., Wallace et al. 1988; Gulev et al. 2002). The LFV of the midlatitude jet can be isolated, for example, by zonal averaging of the atmospheric zonal wind, which effectively filters out synoptic waves. This statistical decomposition can then be substituted in the dynamical equations and used to analyze the mechanisms of the atmospheric LFV (Hartmann 1995; Lorenz and Hartmann 2001, 2003; Barnes and Hartmann 2010; Cai et al. 2007). Other spatial filters have also been widely used to define planetary and synoptic-scale processes (Lupo and Smith 1995; Hoskins and Hodges 2002). Yet another objective filtering technique routinely used in atmospheric studies relies on the empirical orthogonal function (EOF) analysis, which reveals dominant low-frequency flow patterns that may not, however, necessarily be directly interpretable dynamically (Monahan et al. 2009).

An alternative, system-centered method of looking at the atmospheric general circulation is to recognize traveling synoptic eddies as natural building blocks of the atmospheric variability and perform their detection and tracking (e.g., Hoskins and Hodges 2002; Zolina and Gulev 2002; Rudeva and Gulev 2007, 2011; Raible et al. 2008; Dacre and Gray 2009). The atmospheric LFV can then be diagnosed through analyzing the changes in eddy life cycles and paths. This approach reduces the amount of original raw data while maintaining contact with the actual flow field, without invoking any type of time or spatial filtering. For the subsequent analysis of this reduced data, one may apply, for example, the ensemble averaging over many realizations of eddy life cycles (Rudeva and Gulev 2007, 2011) or compositing conditioned on a large-scale flow (Ayrault et al. 1995). Hoskins and Hodges (2002) combined the Eulerian and system-tracking approaches to study the Northern Hemisphere’s wintertime storm tracks. They demonstrated how synthesis of different types of information available through each methodology results in a more complete description of and possibly new insights into the storm-track variability.

The feature that makes the midlatitude atmospheric variability problem difficult to tackle (and thus so interesting!) is that the interaction between high-frequency and low-frequency transients may yield the entirely new phenomena such as the eddy-driven jet (Williams et al. 2007; Vallis and Gerber 2008). The eddy-driven jet arises because of eddy momentum fluxes that rectify the eastward zonal flow at the latitude of eddy generation; this phenomenon was initially described in terms of the “negative eddy viscosity” (Lorenz 1967; Starr 1968). While the synoptic eddies grow via baroclinic instability of large-scale flow, the midlatitude zonal-jet climatology is largely the result of the fundamentally nonlinear eddy–mean flow interaction, making it difficult to identify the causality in the eddy–jet system.

The same eddy–mean flow dualism is manifested in various theories of atmospheric LFV. In one of the existing paradigms, the LFV arises via deterministic, linear, or nonlinear large-scale modes (Frederiksen 1983; Simmons et al. 1983; Legras and Ghil 1985) possibly energized by relatively high-frequency synoptic-eddy forcing (Farrell and Ioannou 1993, 1995; Kravtsov et al. 2003). It is generally accepted that the large-scale flow anomalies do organize the structure of the eddy forcing that excited these anomalies in the first place in a way to reinforce and maintain themselves (Branstator 1992, 1995; Robinson 1994, 1996, 2000, 2006; Cuff and Cai 1995; Feldstein and Lee 1998; Orlanski 1998; Lorenz and Hartmann 2001, 2003; Gerber and Vallis 2007; Son et al. 2008; Barnes and Hartmann 2010), although the mechanisms of this so-called synoptic-eddy feedback are still a matter of debate. Kidston et al. (2010) provided a comprehensive summary of the possible synoptic-eddy feedback dynamics. Barotropic explanations (Hartmann 2000; Limpasuvan and Hartmann 2000) involve response of the late stages of eddy life cycles to the changes in the meridional shear of the westerly jet that accompanies changes in the jet latitude, giving rise to the preferred anticyclonic (LC1) or cyclonic (LC2) wave breaking (Thorncroft et al. 1993; Hartmann 1995) and the resulting maintenance of the displaced jet. On the other hand, baroclinic explanations concentrate on the enhanced eddy generation at the core of the variable eddy-driven jet (Held 1975; Robinson 1994, 1996, 2000, 2006; Gerber and Vallis 2007; Kidston et al. 2010).

The above theories implicitly regard the jet as a primary dynamical entity varying on long time scales and generating secondary high-frequency synoptic eddies, whose collective effect is to further modify the jet. In an alternative paradigm the eddies are, in a sense, primary. Termination of the eddy life cycles involves loss of upper-level support and synoptic wave breaking in the upper troposphere; the remnants of these breaking waves downstream comprise the low-frequency anomalies such as the North Atlantic Oscillation (NAO) (Benedict et al. 2004; Franzke et al. 2004; Vallis and Gerber 2008). This view of the LFV as the one stemming from the dissipation of synoptic eddies was implicit in the study of Löptien and Ruprecht (2005), who used an Eulerian approach to define the time-evolving eddy field and showed that low-frequency variations in the eddy-field distribution alone produce NAO-type variability (e.g., Branstator 2002), consistent with earlier conclusions by Madden (1976). In a nutshell, the midlatitude jet and its variability represents, to a large extent, changes in the characteristics of
the groups of propagating synoptic eddies whose zonal (or temporal) averaging forms the jet; the latter jet, however, may not necessarily be detectable in individual time moments. These arguments also paraphrase the conclusion of Vallis and Gerber (2008) that “the NAO is the variability of an Atlantic storm track.” In contrast, Stephenson et al. (2000) detected long-range correlations in the wintertime NAO index that are inconsistent with the picture of interannual NAO variability being purely the result of sampling fluctuations in the number and intensity of the North Atlantic synoptic eddies, suggesting a role for large-scale dynamics in the observed LFV.

The purpose of our study is to further explore and reconcile the above two paradigms of the coexistence and symbiosis between synoptic eddies and mid-latitude jet in the simplest dynamical configuration of a two-layer $\beta$-plane quasigeostrophic channel model (Kravtsov et al. 2005). We will study the “anatomy” of the low-frequency flow in terms of eddy tracking and characteristics of these eddies, address the following major questions, and discuss our answers in terms of the related studies and existing theories of the atmospheric LFV:

1) To what extent do the variations in synoptic-eddy life cycle characteristics and paths represent the full atmospheric variability simulated by the model?

2) What are the relative contributions of the case-to-case variability in eddy strength, eddy radial asymmetries, and the eddy paths in the fraction of the model’s variability associated with synoptic eddies?

3) How do eddy life cycles depend on the phase of the simulated LFV? In particular, are the meridionally displaced and anomalously persistent jet states characterized by enhanced generation of synoptic eddies relative to the eddy generation during (less persistent) episodes when the jet is near its climatological position? What are the major differences in eddy paths between the climatological and shifted jet states?

4) Does random kinematic redistribution of synoptic eddies fully explain jet shifts? Is there evidence of a larger-scale feedback between eddies and a mean flow?

To address these fundamental questions, we chose to use an atmospheric model with idealized geometry and simplified dynamics. This will allow us to focus on a minimal subset of potentially important processes for midlatitude LFV, while exploring a wide sector of the model’s parameter space. We follow Hoskins and Hodges (2002) in combining the Eulerian and feature-tracking methodologies to identify the model’s LFV, but taking their approach one step further by developing a technique to unify and relate the results obtained via each methodology. This technique for studying the kinematics of eddy–mean flow interaction is model independent and can be applied to other observed and model-generated datasets; its novel centerpiece is the setup of a surrogate synoptic-eddy field via launching round cyclones and anticyclones evolving according to their composite life cycles along the actual eddy tracks simulated by the model in each parameter setting.

The paper is organized as follows. The model, experimental setup, and analysis methodology are presented in section 2. We will describe the model’s LFV in terms of the EOFs in section 3 and use cyclone/anticyclone tracking to quantify characteristics of the synoptic eddies during different phases of the LFV (section 4). We summarize and discuss our results in section 5.

2. Model setup and analysis methodology

a. Model setup and the strategy of experiments

We used the two-layer $\beta$-plane quasigeostrophic channel model described in detail in Kravtsov et al. (2005), augmented by the inclusion of wavenumber-2 sinusoidal topography (outlined by white contours in Fig. 1) with maximum height $h$ measured from trough to crest (values we used will be listed below). The model has the horizontal resolution of 200 km and spans the entire hemisphere, with 128 grid points in the longitudinal direction and 49 points in latitudinal direction. It is forced, at the interface between the lower and upper model layer, with the heat flux function $F$ of the form

$$F = \frac{Q_0}{4} [1 + Q_2 P_2(\chi)][1 - a_0 - a_2 P_2(\chi)] - A - BT_a,$$

(1)

where $Q_0$ is the solar constant (values used will also be listed below), $a_0 = 0.3$ is the net Earth albedo, $P_2(\chi)$ is the second Legendre polynomial expressed in terms of the sine of the latitude $\chi$, $Q_2 = -0.482$ and $a_2 = 0.14$ are the dimensionless parameters governing the equator-to-pole gradient of the radiative forcing, and $A = 190 \text{ W m}^{-2}$ and $B = 1.63 \text{ W m}^{-2} \text{C}^{-1}$ are the parameters determining the intensity of infrared radiation emission to space. The quantity $T_a$ is the atmospheric temperature linearly related to the model’s baroclinic streamfunction. The thickness of the model’s lower layer was characterized by the surface drag coefficient $k$. The value of $k$ corresponded to the Ekman-layer depth of 450 m, or the damping time scale of about 11 days for the barotropic mode.

To explore a wide range of possible eddy–mean flow interaction regimes defined via the set of major questions
posed in section 1, we varied $Q_0$, $h$, as well as the model’s Rossby radius of deformation $R_d$ (see Table 1). The major expected effect of increasing $Q_0$ in the present idealized model is to increase equator-to-pole thermal contrast and thus intensify the circulation, increasing $h$ enhances zonal symmetry of the simulated flow, while $R_d$ affects, among other things, the spatial scales and propagation speeds of dominant waves in the model.

We performed eighteen 10-yr-long experiments with all possible combinations of $Q_0$, $h$, and $R_d$ values listed in Table 1. All experiments involved 1-yr spinup from rest and subsequent 10-yr integrations with the model fields stored every 6 h of model time. The selected range of the parameter values produced a fairly broad range of model behaviors. However, since the results of the warmed (W) suite of experiments are very similar to those from the corresponding members of the present-day (P) experiments, with only slightly more vigorous variability in the W experiments, we will illustrate the majority of our results by visualizing the outcome of the P experiment with median values of $h$ and $R_d$ parameters: $h = 2$ km (H2) and $R_d = 600$ km (R6), which is hereafter abbreviated, according to the Table 1, as the experiment P_H2_R6.

The model’s lower-layer climatology is characterized by a wave-2 midlatitude eastward jet (Fig. 1a), the jet’s

![Image](https://example.com/image1.jpg)

**Fig. 1.** (a),(c) Time-mean and (b),(d) variance of the lower-layer streamfunction in the P_H2_R6 run (see Table 1) for (a),(b) the full streamfunction field generated by the QG model and (c),(d) a synthetic streamfunction field obtained via eddy tracking (see text for details). White contours show locations of the mountains. The streamfunction fields are dimensionless, normalized by $L^2/T$ using the length scale of $L = 200$ km (grid size) and velocity scale of $10 \text{ m s}^{-1}$ (hence the time scale of $T = 200000/10 = 20000\text{s} \approx 5.5\text{h}$).

### TABLE 1. Model parameters and acronyms of the experiments.

The latter acronyms are formed by combining the notations referring to individual parameter values, as listed in this table. For example, the experiment with $Q_0 = 1355$ W m$^{-2}$ (P), $h = 2$ km (H2), and $R_d = 600$ km (R6) is denoted as P_H2_R6.

<table>
<thead>
<tr>
<th>Model parameter</th>
<th>Parameter value</th>
<th>Expt notation</th>
</tr>
</thead>
<tbody>
<tr>
<td>$Q_0$ (W m$^{-2}$)</td>
<td>1355</td>
<td>Present (P)</td>
</tr>
<tr>
<td></td>
<td>1355 $\times$ 1.05</td>
<td>Warmed (W)</td>
</tr>
<tr>
<td>$h$ (km)</td>
<td>0</td>
<td>H0</td>
</tr>
<tr>
<td></td>
<td>2</td>
<td>H2</td>
</tr>
<tr>
<td></td>
<td>4</td>
<td>H4</td>
</tr>
<tr>
<td>$R_d$ (km)</td>
<td>400</td>
<td>R4</td>
</tr>
<tr>
<td></td>
<td>600</td>
<td>R6</td>
</tr>
<tr>
<td></td>
<td>800</td>
<td>R8</td>
</tr>
</tbody>
</table>
zonal modulation being in phase with topography. The streamfunction variance (Fig. 1b) is also at maximum in the middle latitudes and also has a wave-2 zonal modulation, which is, however, nearly out of phase with topography. The climatology and variance of the surrogate streamfunction field constructed using synoptic-eddy tracks with actual, full-model-generated paths but composite-mean eddy life cycles (see sections 2b and 2c) reveal the patterns that are nearly identical to those from the full-model simulation (cf. Figs. 1c,d and Figs. 1a,b). According to Figs. 1c and 1d, the simulated surrogate streamfunction variance is slightly higher, and its wave-2 zonal modulation is slightly more pronounced than in the full streamfunction field shown in Figs. 1a and 1b.

b. Eddy tracking

Cyclone activity identified in reanalysis datasets by various tracking algorithms (Gulev et al. 2001; Zolina and Gulev 2002; Hoskins and Hodges 2002; Rudeva and Gulev 2007; Raible et al. 2008; Neu et al. 2013) reveals a wide range of behaviors, especially in regions of complex orography or in the areas where atmospheric transients slow down, such as the Icelandic low region. There are a few alternative methods of cyclone tracking in observations: some methods analyze surface pressure fields, some methods use vorticity, and some methods apply combined approaches. Recently, a large international effort, the Intercomparison of Mid-Latitude Storm Diagnostics (IMILAST) project (Neu et al. 2013) was undertaken to quantify the uncertainties of storm-tracking algorithms. The results showed large quantitative differences between the tracking methodologies, but they also clearly showed that these differences cannot be attributed to the usage of vorticity or mean sea level pressure to track cyclones, or to preprocessing via large-scale spatial filters (Sinclair 1997; Anderson et al. 2003). Instead, the tracking-scheme uncertainties and large scheme-to-scheme differences turned out to be associated with contributions from relatively shallow, short-lived (mostly continental) cyclones, multicenter stationary depressions, and cyclogenesis over the complicated orography. The eddies in our idealized model naturally exhibit a less diverse behavior, and sources of the uncertainties in eddy tracking mentioned above are either nonexistent or play at most a minor role.

Our methodology most closely follows one of the tracking schemes tested in Neu et al. (2013), namely, the one that identifies eddies in surface pressure fields, whose proxy in our model is represented by the lower-layer streamfunction. To perform tracking of eddies in our simulations we first identified, for each time step of the model output, closed contours of the lower-layer streamfunction’s deviation from the basin-mean value (see Fig. 2); note that we did not count the “circum-global” closed streamlines that only had a single intersection with each of the latitudinal “boundaries” of the model’s reentrant channel. We used 100 contours spanning the range between the minimum (maximum) of the streamfunction anomaly and zero to identify cyclones (anticyclones). For each closed contour, we computed the contour’s center of mass (the average coordinate of the points belonging to the contour), as well as the area within this contour. Since the centers of mass of the closed contours for a single cyclone or anticyclone are near one another, we used a hierarchical clustering algorithm applied to the data matrix of the centers of mass to identify contour groups that defined a single eddy. Within each such group, the center of mass of the contour with the smallest area was identified with the eddy-center location \((x_0, y_0)\). The bogus eddy groups represented by the closed contours that embed several eddies were dealt with by discarding the minimal-area contours whose area values were far outside of the range defined by true single-eddy structures.

To track the eddy evolution through time, we used a minimum-distance clustering algorithm applied to the 14600 × 3 matrix of 10-yr-long, 6-hourly time series of \([x_0(t), y_0(t), t]\). In this algorithm, the distance was defined as the normal Euclidean distance in the periodic channel for any two points separated by one time unit (6 h);
otherwise, the distance was set to infinity. Hence, in the hierarchical cluster tree, the children clusters were always characterized by continuous time. The progression of \([x_0(t), y_0(t)]\) through this continuous time segment defined the eddy path. An example of cyclone/anticyclone tracking is shown in Fig. 2.

For each cyclone (anticyclone), we computed the eddy intensity \(I_0\) as the minimum (maximum) streamfunction anomaly, and the eddy effective radius \(R_p\). The radius \(r\) of each contour was computed as \(r = \sqrt{A/\pi}\), where \(A\) is the area of the contour. Within each group of contours, we then identified the contour with median intensity and set \(R_0 = r_{\text{median}}\). We also defined, following Rudeva and Gulev (2007), the eddy asymmetry ratio \(\eta = r_{\text{min}}/r_{\text{max}}\) where \(r_{\text{min}}\) is the minimum distance from the center to the median-intensity contour and \(r_{\text{max}}\) is the maximum distance. Each eddy track was thus characterized by the set of time-varying characteristics \((x_0, y_0, I_0, R_0, \eta)\) that were defined during the eddy lifetime. The latter was found as the time interval between the first appearance and eventual decay of the closed contour associated with this eddy.

c. Construction of the surrogate streamfunction field

The eddy life cycles in our model exhibit a robust self-similarity, with eddies that live longer, achieving larger maximum sizes and intensities (not shown); this is also in agreement with the observed cyclone life cycles (Simmonds 2000; Rudeva and Gulev 2007, 2011; Schneidereit et al. 2010). To construct a surrogate lower-layer streamfunction field, we first computed—separately for all cyclone and anticyclone tracks with a given lifetime (using daily bins)—the ensemble-mean time series of the eddy intensity \(I_0\) and effective radius \(R_0\).

Then, we replaced the actual time series of \(I_0\) and \(R_0\) for each track with this lifetime by the composite-mean time series \(\bar{I}_0\) and \(\bar{R}_0\), while keeping the original, full-model time series of the eddy position \((x_0, y_0)\). We repeated this procedure across the entire range of eddy lifetimes simulated by the model. The majority of the eddy tracks had typical lifetimes of less than 10 days, and within this group, each composite was based on over a few hundred tracks. The eddy groups with longer lifetimes of 10–15 days typically contained at least a few dozen tracks, while the eddies with longer lifetimes were rare.

The contribution of each eddy centered, at a given time, at \((x_0, y_0)\) to the surrogate lower-layer streamfunction field was then approximated by the exponential function

\[
F(r) = c\bar{I}_0 \exp(-r/\bar{R}_0),
\]

where \(r\) is the radial distance from \((x_0, y_0)\) and \(c\) is the constant coefficient computed so that the synthetic streamfunction’s climatology best matches the full-model climatology; typical values of \(c\) are around 2. The synthetic streamfunction field for each time was obtained by adding streamfunction contributions (2) from all cyclones and anticyclones present at that time; this gives us the 10-yr-long, 6-hourly sampled time series of the synthetic (or reconstructed) streamfunction.

Note that while eddies in the reconstructed flow have been symmetrized and their sizes and strengths were determined from composites, their locations fully retain the variability of the full model. In other words, the reconstruction retains the full variability in eddy tracks in the model but has discarded case-to-case variability in eddy strength and all eddy asymmetries. Thus, the synoptic eddy field obtained in this way represents the variability associated with the distribution of synoptic systems in the model, since the track-to-track variability in the eddy life cycles with a given lifetime, as well as the small-scale variability due to eddy asymmetries, have been effectively filtered out. The essence of the above procedure is analogous to that of the Eulerian filtering method used by Löptien and Ruprecht (2005), but the present method is arguably advantageous to the latter in being somewhat less ad hoc.

3. EOF analysis of original and surrogate fields

a. Streamfunction analysis

We used EOF analysis (Monahan et al. 2009) to compactly describe dominant modes of variability of the simulated model fields. The EOF patterns and the corresponding principal component (PC) time series for the lower- and upper-level streamfunction fields are well correlated (see Fig. 3 for P_H2_R6 run; the EOF patterns for other runs are qualitatively similar). The leading EOF in Fig. 3 is predominantly zonally uniform and reflects the north–south migrations of the climatological jet axis; we will therefore refer to this mode as the jet-shifting (JS) mode. In addition to the JS mode, the simulated atmospheric variability is characterized by propagating waves, which appear in the EOF analysis as pairs of modes with a given wavenumber: EOFs 2–3 represent wave 3 (W3); EOFs 4–5 represent wave 4 (W4), and EOFs 9–10 (EOF 10 is not shown) represent wave 5 (W5). The remaining EOFs describe the stationary jet-intensification mode (EOF 6) and wave-2 longitudinal modulations of the wave-4 train (EOFs 7–8).

The latter statement about the EOFs 7–8 complementing the wave-4 variability described by EOFs 4–5 was based on the structure of the autocorrelation function (ACF) of the PCs corresponding to these modes (see Fig. 4). In particular, the ACF of PCs 7–8
exhibits a decay modulated by oscillatory wiggles with the time scale of about 5 days, which corresponds well to the periodicity evident in the ACFs of PCs 4–5. The ACFs for the stationary modes EOF 1 and EOF 4 are characterized by slow monotonic decay, while the wave modes (EOFs 2–3, EOFs 5–6, EOFs 9–10) exhibit damped oscillations. In general, the shorter time scale modes account for a progressively smaller fraction of variance in the EOF spectrum, but there is no clear time scale separation between all these modes. The JS, W3, W4, and W5 modes are robust and exist throughout the range of model parameters explored, albeit their relative magnitudes vary.

b. Analysis of the surrogate streamfunction

Figure 5 shows the EOFs of the lower-layer surrogate streamfunction, which represents the distribution of the synoptic systems (section 2c). The spatial correlations between the corresponding EOFs and time correlations between the corresponding PCs are high, and the explained variances are similar for the leading JS, W3, and W4 modes of the actual and surrogate fields (cf. Figs. 3, 5). Similar time dependences of the actual and surrogate propagating modes are also confirmed by nearly identical structures of their ACFs (Fig. 4). The trailing W5 EOF modes, as well as larger-scale modes evident for EOFs 7–8 (Fig. 3) are also present in the surrogate streamfunction field (Fig. 5), but account for a relatively smaller fraction of variance (not shown). All in all, it can be concluded that the surrogate streamfunction field composed of the synoptic eddies contains a great deal of information about the full streamfunction field, including its time-mean, LFV, and leading propagating modes.
c. EOF analysis of zonal wind

We used the EOF analysis of the zonal-mean zonal wind $\langle u \rangle$ to characterize variability of the leading JS mode, since the latter has a pronounced zonal-mean component. The results for all P runs (Figs. 6 and 7) are, once again, analogous to those for the W runs (not shown).

The EOF patterns in Fig. 6 are dimensional (m s$^{-1}$), obtained by regressing the $\langle u \rangle$ time series onto the leading normalized PC of this field; hence, the amplitude of each such EOF describes the intensity of the JS mode in a given run, while the meridional extent of the jet shifts is measured by the distance between major crest and trough of the zonal-wind EOF pattern. Both the intensity and spatial scale of the JS mode are practically insensitive to $h$ and have similar values for different $h$ (given $R_d$). The dependence on $R_d$ is more substantial, with the jet shifts growing in magnitude and intensity with increasing $R_d$. The $\langle u \rangle$ EOFs of the surrogate fields trace the structure of the actual $\langle u \rangle$ EOFs quite well throughout the central part of the basin, being somewhat squished toward the channel axis near the boundaries, likely because of failure of the round-eddy approximation. Despite that, the corresponding leading zonal-wind PCs of the actual and surrogate fields are well correlated with each other, while the leading zonal-wind PCs of the full lower-layer $\langle u \rangle$ are well correlated with the leading PCs of the full lower-layer streamfunction (such as the PC for the leading EOF shown in Fig. 3).

The similarity between the zonal-mean zonal wind PCs of the full and synoptic-eddy reconstructed flow...
fields is further confirmed by nearly identical structure of the zonal-wind-based JS-mode ACFs (Fig. 7). Note that the time scale of the JS mode clearly grows with $R_d$ and shortens with increasing $h$. The time scale of the JS mode is likely to be partly controlled by the interactions with the eddy field, which, in turn, depends on the background climatological state (e.g., Son et al. 2008).

d. Summary of EOF analyses

The model’s LFV is dominated by the JS mode, whose intensity and persistence are likely to depend on interactions with propagating wave modes. The latter modes are orthogonalized and ordered manifestations of the synoptic eddies—cycloons and anticyclones—identified by tracking. In particular, the surrogate flow fields representing the distribution of the synoptic systems and composed of composite eddy life cycles launched along the actual simulated cyclooon/anticyclone tracks have the same LFV and leading propagating modes as the full model fields.

4. Synoptic-eddy tracks and LFV of the jet–eddy system

a. Composite eddy life cycles

It turns out that the evolution of the eddy size and intensity for tracks with different lifetimes is characterized by the same shape in the stretched time coordinate system nondimensionalized by the eddy lifetime (Rudeva and Gulev 2007). This allows one to compute meaningful composite distributions of the eddy characteristics during various stages of the dimensionless eddy lifetime spanning the range from 0 to 1. Figure 8...
shows an example for the composite-mean lower- and upper-layer streamfunction fields in the coordinate system associated with a moving cyclone for the simulation P_H2_R6. A shallow minimum of the lower-layer streamfunction at initial stage of the cyclone development deepens as the cyclone grows in size and intensity and reaches the maximum strength at the midpoint of its life cycle, after which the size and intensity decrease during the cyclone decay stage. The upper-layer streamfunction exhibits a trough westward of the lower-layer streamfunction minimum during the cyclone growth stage; this tilt decreases during the formation of the upper-layer streamfunction’s closed contour during the intermediate stage and essentially vanishes in the final stage of the life cycle (cf. Schneidereit et al. 2010). The composite evolution of the lower-level vorticity field (not shown) closely resembles the streamfunction composites (albeit, of course, with a minus sign), thus demonstrating in part that our tracking results are insensitive to tracking methodology.

We also display the evolution of probability density functions (PDFs) of $R_0$, $I_0$, and $\eta$ (section 2b) through the dimensionless lifetime (Fig. 9). The cyclone size $R_0$ (Figs. 9a,d) grows in the first half of the cycle and decays in the second half, with a larger spread of $R_0$ at the later stage of the life cycle ($0.6 < t/T < 0.8$) relative to the early stage ($0.2 < t/T < 0.4$). The PDFs of intensity (Figs. 9b,e) and of the asymmetry ratio (Figs. 9c,f) exhibit a more symmetric behavior. Overall, despite the model’s simplicity, the distributions of the eddy characteristics
during dimensionless eddy lifetime are qualitatively and quantitatively similar to the observed eddy life cycles (Simmonds 2000; Rudeva and Gulev 2007; Schneidereit et al. 2010).

Close similarity of the life cycles for cyclones and anticyclones (Figs. 9a–c and 9d–f) in our model may not be entirely surprising and is due to intrinsic symmetry of approximate quasigeostrophic (QG) equations and the nearly antisymmetric, about the channel axis, model's radiative forcing. In reality, the non-QG and diabatic processes result in considerable differences between life cycles of cyclones and anticyclones (Hoskins and Hodges 2002). The major and only difference between these life cycles in our simplified model, however, is in the location of their tracks. The PDFs of the cyclone and anticyclone locations at the beginning and at the end of the track for the P_H4_R6 case (Fig. 10a) show that the cyclone generation has two major locations on the leeward side of each model mountain and two secondary locations just westward of the mountain. The cyclone decay sites (Fig. 10b) are located just north of the secondary generation sites. The structure for the anticyclones is similar, but the generation/decay locations are shifted with respect to those for cyclones. The anticyclone primary generation occurs westward of the mountains and the secondary generation occurs on the leeward side of the mountains (Fig. 10c), with the dissipation sites located southward of the secondary generation sites (Fig. 10d). Thus, most of the cyclone tracks generated at the leeward side of the mountain travel northeastward to their dissipation site. Some of these cyclones have more zonal tracks and do not disappear, reemerging at the secondary formation sites from the remnants of the dissipating cyclones formed at primary generation location (cf. Mailier et al. 2006). Zonal asymmetries in the eddy generation/dissipation sites
FIG. 8. Cyclone composite evolution over dimensionless time (with 0 corresponding to the beginning of the cyclone track and 1 corresponding to the end of the cyclone track) for the P_H2_R6 run. The composites are computed in the coordinate system centered at the current cyclone location (dx/dy = 200 km) and show both lower-layer (shading) and upper-layer (contours) streamfunction (positive contours thick).
naturally disappear in the zonally symmetric cases with \( h = 0 \) (not shown). These cases are characterized by the zonally uniform strips of cyclone (anticyclone) generation poleward (equatorward) of the zonal-mean zonal wind maximum and dissipation sites north (south) of the generation sites.

**b. Changes in eddy life cycles between different phases of LFV**

We now examine the characteristics of eddy life cycles over the course of the dominant jet-shifting LFV defined via the leading EOF of \( \langle u \rangle \) (Fig. 6). For each simulation, we tag the days during which the leading PC of \( \langle u \rangle \) exceeds the threshold of +1 standard deviation as belonging to the positive JS (JS+) phase of the LFV, while all the days during which this leading PC has values below the threshold of −1 standard deviation as those of the negative JS (JS−) phase of the LFV. We then compute composite characteristics of the eddies for JS+ and JS− phases of the LFV and compare them with each other, as well as with the climatological eddy characteristics composited over all tracks.

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**FIG. 9.** The evolution, over dimensionless time normalized by the eddy’s lifetime, of the PDFs of the (a),(d) eddy effective radius, (b),(e) normalized eddy intensity, and (c),(f) eddy asymmetry ratio for (a)–(c) cyclones and (d)–(f) anticyclones. The results are shown for the P_H2_R6 run.
Figure 11 shows the climatological, JS⁺, and JS⁻ composites of cyclone and anticyclone tracks computed over dimensionless eddy lifetime, along with the corresponding composites of ⟨u⟩ over the course of the model’s LFV for all P simulations. The results for the W case are analogous. The time-mean zonal-mean jet becomes slightly wider with increasing $R_d$ and is insensitive to $h$ (with a given $R_d$), while the jet’s intensity and latitudinal locations change very little across the whole range of the parameters explored. The JS mode is characterized by a slight intensification of the jet in both JS⁺ and JS⁻ phases. The shifted jets in the R8 cases are also slightly narrower than the climatological jet. The intensity of the JS mode visualized through the JS⁺ and JS⁻ jet composites in Fig. 11 corroborates sensitivities evident in Fig. 6: the jet shifts and intensifications both become more pronounced with increasing $R_d$, while the sensitivity to $h$ (given $R_d$) is small.

The cyclone (anticyclone) tracks are located to the north (south) of the zonal-jet axis and deflect farther to the north (south) during their evolution. This is true for both the climatological (all tracks) and JS⁺/JS⁻ composites. In the first approximation, the eddy tracks shift in latitude along with the meridional jet shifts. However, the all-track-based cyclone and anticyclone composites (Fig. 11) are more symmetric about the jet axis with respect to each other than the JS⁺ and JS⁻ cyclone/anticyclone tracks. The JS⁺ state is characterized by the longer and more zonal (less curved) cyclone tracks compared with shorter and strongly curved, at the late stages of the life cycle, anticyclone tracks. In the JS⁻ state, these relative cyclone/anticyclone track characteristics are reversed. This behavior is most apparent in the R8 cases, in which the cyclone/anticyclone tracks are the longest, but is also noticeable for smaller $R_d$ corresponding to the shorter eddy tracks.

Figures 12 and 13 show the dependencies of eddy lifetimes and maximum intensities on model parameters, for both their climatological values and during the JS⁺ and JS⁻ states of the LFV. Maximum intensity and effective eddy radius computations (the latter not shown) used the surrogate eddy fields, with filtered out
Fig. 11. Zonal-wind variability and composite cyclone/anticyclone tracks for the present-day runs. Different panels correspond to the runs with different values of \( h \) and \( R_d \) (as indicated in each panel). The \( y \)-axis in each panel corresponds to the \( y \) coordinate (in grid points) across the model channel. The top \( x \)-axis labels measure the lower-layer zonal-mean zonal velocity: time-mean velocity (black lines), velocity composite over the JS\(_1\) states with the leading zonal-mean zonal wind PC (corresponding to the EOFs in Fig. 6) exceeding \(+1\) of its standard deviation (red lines), and velocity composite for the JS\(_2\) states with this PC’s value below \(-1\) of its standard deviation (blue lines). The bottom \( x \)-axis labels measure the composite zonal displacement (in grid points) of the eddy relative to its initial location; these values correspond to the composite eddy trajectories for cyclones (solid lines with dots) and anticyclones (dashed lines with dots). The trajectories are color-coded in the same way as the zonal-mean zonal wind composites for the time-mean (black), JS\(_+\) (red), and JS\(_-\) (blue) states.
FIG. 12. Mean eddy lifetimes for the present-day runs with different values of $h$ (shown in the title of each panel) as a function of $R_d$ ($x$ axis of each panel). The averaging is done over all tracks (black lines) as well as tracks during JS+ (red lines) and JS− (blue lines) states of the zonal-mean jet (see Fig. 11). The 95% confidence error bars represent standard deviations of these quantities multiplied by $\sqrt{4/N}$, where $N$ is the effective number of tracks within a given category (all tracks for time mean, or JS+/JS− tracks). Results are for (a),(c),(e) cyclones and (b),(d),(f) anticyclones.
variability in these quantities among the tracks with a given lifetime (section 2c). This is justified since the full and surrogate eddy fields have nearly identical time means (Fig. 1) and very similar leading EOFs (Figs. 3–5).

Eddy lifetimes, intensities, and radii (not shown) all increase with $R_d$, which is consistent with the larger sizes and intensities of the longer-lived eddies. Furthermore, the eddies with lifetimes exceeding the climatological
lifetime populate the JS+/JS− phases of the LFV. This is indicative of the enhanced persistence of the jet-shifted states. The JS+/JS− eddies are also slightly more intense (Fig. 13) and have a slightly larger effective radii (not shown) compared to the climatological eddy sizes and intensities. All these results pertain to the W runs as well.

In summary, the persistent latitudinal shifts and intensifications of the zonal-mean jet in our model are accompanied by the coordinated shifts and increase in the mean lifetime of the synoptic eddies. The northward jet shifts are characterized by more zonal cyclone tracks and shorter anticyclone tracks curving southward and becoming nearly stationary at the latest stages of the life cycle, while the southward jet shifts exhibit longer and more zonal anticyclone tracks and relatively short cyclone tracks curved to the north and also slowing down in the end of their life cycle. This asymmetry may be related to the primary mechanism of the synoptic-eddy feedback in the present model (see section 5b for more discussion).

c. Are the jet shifts a mere consequence of the synoptic systems’ redistributions?

Our results demonstrate intimate connections between the distribution of the synoptic systems and the model’s dominant LFV composed of the predominantly zonally uniform changes of the midlatitude jet. However, one can argue, similarly to Löptien and Ruprecht (2005), that the midlatitude jet and its LFV are mere consequences of the distribution of synoptic systems and its random low-frequency changes over time. The counterarguments in favor of an alternative interpretation may consider thought experiments involving the spinup of the climatological jet. For example, when the model is launched from the state of rest, the large-scale thermal forcing induces, according to the thermal-wind balance, a strong wind shear between the lower and upper layers in the jet. This configuration is unstable, and synoptic eddies—which arise via baroclinic instability of the large-scale flow—interact with this flow and modify it to produce a relatively narrow and intense midlatitude jet. Can this line of thought be extended to the LFV of the turbulent midlatitude jet? Specifically, is there evidence that the LFV of the jet in the present model is more than a manifestation of random variations in the eddy statistics? Does LFV involve larger-scale processes characterized by longer-range (relative to eddy size) spatial correlations that contribute to organizing eddy tracks?

To address these questions, we analyzed the time evolution of the zonal-mean jet in relation to the storm-track latitude. We defined the instantaneous storm-track position at a given time by averaging the latitudinal coordinates of all simultaneously existing cyclones and anticyclones; the resulting storm-track latitude time series was centered and normalized to unit variance. We also treated the leading PC of $\langle u \rangle$ as a proxy for the zonal-jet position. A striking and robust property of the cross correlation between the storm-track and zonal-mean jet position time series so defined is its shift with respect to the zero-lag axis, indicating that the zonal jet persistently leads the storm tracks, as the cross correlations at positive lags are consistently higher than the cross correlations at negative lags [see Figs. 14a,d for the P_H0_R4 and P_H2_R6 runs, respectively, and P and W runs (not shown) produce qualitatively analogous results]. In other words, the zonal-mean jet position is a more skillful statistical predictor for the storm-track position than vice versa. The spectral analysis (Figs. 14b,c and 14e,f) shows that 1) the storm-track latitude has more variance than jet latitude at high frequencies, while their variances at low frequencies are essentially the same (Figs. 14b,e); 2) at high frequencies, the eddies lead the jet, while the reverse is true at low frequencies; and 3) there is a “cross-spectral” gap of low squared coherence between eddies and the jet separating the frequency domains in which the eddies lead/lag the jet; at low frequencies, the squared coherence tends to unity. These properties argue for the zonal-mean jet shifts being dynamically more involved than simply reflecting an artifact of averaging over low-frequency changes in the distribution of the synoptic systems, thus indicating a possible dominant influence of larger-scale processes on the jet’s evolution.

5. Summary and discussion

a. Summary

We performed a series of experiments with an idealized baroclinic channel model of the atmosphere to investigate the association between synoptic-eddy life cycles and atmospheric low-frequency variability (LFV). The simulations over a range of control parameters including the height of the model’s wave-2 topography $h$ and Rossby radius $Rd$ were analyzed in two principally different ways: 1) via empirical orthogonal function (EOF) analysis of the streamfunction and zonal-mean zonal wind fields and 2) by tracking and studying collectively the evolution of individual cyclones and anticyclones in the model.

The central result of our paper, which answers the first two questions raised in the introduction, is that the surrogate synoptic-eddy field constructed by launching the composite-mean round cyclones and anticyclones
along their actual simulated tracks in the model has the same time mean and very similar leading modes of variability compared to the full streamfunction field. Thus, the distribution of synoptic eddies carries a lot of information not only about the propagating high-frequency modes, but also about the LFV.

The leading mode of the LFV in the model is associated with the north–south displacements and simultaneous
Intensification of the zonal-mean jet. Compositing the cyclone and anticyclone tracks over the positive (JS+) and negative (JS−) phases of this jet-shifting variability addresses the third question and reveals that 1) the cyclone/anticyclone tracks shift in latitude along with the jet; 2) the lifetimes of the eddies in the shifted jet states exceed climatological eddy lifetime, indicating enhanced persistence of the jet shifts, while other eddy-track characteristics such as the mean effective eddy radius and maximum intensity do not deviate much from their climatological values; and 3) while the climatological (all track) composites of cyclones and anticyclones in the model exhibit essentially the same life cycles, their JS+ and JS− composites are antisymmetric. In the JS+ state, the cyclone tracks to the north of the shifted jet are more zonally elongated, while the anticyclones southward of the shifted jet become stationary and curve to the south at the end of their life cycle. For the JS− state, this behavior is reversed: the anticyclone tracks to the south of the shifted jet are more zonal, while the cyclones to the north of the shifted jet become stationary and curve to the north by the end of their life cycle.

We addressed, statistically, the introduction’s fourth question about causality between the jet-shifting mode and redistribution of synoptic eddies by showing that the storm-track latitude can be better predicted in the medium-to-long time range using the zonal-mean jet latitude information than the jet latitude using the storm-track latitude. This lead–lag relationship indicates the presence of a baroclinic storm track. We attempted, following Hoskins and Hodges (2002), to reveal a different kinematic perspective on the midlatitude eddy–mean flow interaction problem by synthesizing the traditional Eulerian-based methodology with the feature-tracking analysis of LFV. Our results demonstrate that the synoptic eddies are indeed, as tagged in the introduction, the “building blocks” of the simulated midlatitude flow, and the flow’s LFV is dominated by that in the eddy paths.

The results in Fig. 14 imply that while initial zonal-mean jet displacements largely reflect random kinematic redistribution of eddies, there must exist a low-frequency mechanism organizing large-scale flow changes that preemt redistribution of synoptic eddies at a later time. Hence, the maintenance of jet shifts in the model involves processes with correlations in the zonal direction over spatial scales larger than the size of the eddy. These processes may be linear (Farrell and Ioannou 1993, 1995; Kravtsov et al. 2003; Swanson 2007) or nonlinear, involving interactions either exclusively within the slow manifold of the atmospheric dynamical system (Legras and Ghil 1987) or those between atmospheric processes with different scales (Koo and Ghil 2002; Koo et al. 2003; Kravtsov et al. 2005; Swanson 2007, 2008).

The fact that the jet latitude is a better predictor of the storm track than vice versa may also simply indicate that the zonal flow has a built-in flywheel associated with the synoptic-eddy feedback described by Vallis and Gerber (2008) in terms of the “stickiness” of the zonal-mean flow. Our composite analysis indicates that the entire composite tracks of the synoptic eddies shift with the jet, so the latitudes of both eddy generation and synoptic-wave breaking follow the jet. The eddy life cycles in the shifted jet states are very similar to the climatological life cycles, except for the asymmetry between the “shifted” cyclone and anticyclone tracks related to the preferred mode of wave breaking. One may thus argue that the shifted jet states are more persistent (or more “sticky”) and synoptic-eddy lifetimes there are longer due mainly to the barotropic mechanism (Hartmann 2000; Limpasuvan and Hartmann 2000; Barnes and Hartmann 2010), in which enhanced anticyclonic wave breaking (in our results, anticyclone tracks anomalously curving southward) reinforces poleward jet shifts, while cyclonic wave breaking (cyclone tracks anomalously curving northward) reinforces equatorward jet shifts.

Our idealized model is but a metaphor for the real atmosphere. While seemingly reproducing some aspects of the atmospheric behavior, the model is clearly at odds with some of the results obtained in more advanced models, which are not at all captured in our sensitivity experiments. Examples of such phenomena include the dependence of the jet latitude on the eddy length scale (Kidston et al. 2011) or enhanced poleward deflection (Orlanski 1998) of the midlatitude storm tracks, which may qualify as one of the most remarkable features of simulated warmer climates of the future (Löptien et al. 2008; Ulbrich et al. 2009; Woollings and Blackburn 2012); this signal is also present in the instrumental climate record (Bender et al. 2012). While applicability of our model is limited, our analysis methodology is more versatile and can be used in conjunction with the output from GCMs and observations to gain a complementary perspective on the covariability of midlatitude jet stream and storm tracks.
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