Diabatic Damping of Zonal Index Variations
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ABSTRACT

Zonal index variations, or north–south shifts of the midlatitude jet, are the dominant mode of zonal wind variability in the Southern Hemisphere. Previous studies have shown that synoptic-time-scale eddy momentum flux provides a positive feedback and acts to increase the persistence and low-frequency variance of the zonal index. However, the impact of diabatic heating due to the precipitation associated with these eddies has not been investigated. In this study, regression analyses have been conducted to demonstrate that a robust precipitation anomaly can be found to accompany the jet and eddy momentum flux anomalies associated with a poleward shift of the jet, with enhanced precipitation on the poleward flank of the jet and reduced precipitation on the equatorward flank. Diabatic heating associated with such a precipitation anomaly is expected to reduce the temperature gradient across the jet anomaly, thus decreasing eddy generation and damping the anomaly. This expectation is confirmed by three sets of mechanistic model experiments, using three different ways to mimic the impact of moist heating in a dry model. Results of this study suggest that diabatic heating provides a negative feedback to zonal index variations, partially offsetting the positive feedback provided by eddy momentum flux. These results could partially explain why zonal index variations have been found to be very persistent in dry mechanistic model experiments since this negative diabatic feedback is absent in dry models. These results suggest that these models may be overly sensitive to climate forcings that produce a jet shift response.

1. Introduction

Numerous previous studies have shown that the dominant mode of low-frequency variability of the zonal wind in the Southern Hemisphere (SH) is an approximately equivalent barotropic dipole with maximum anomalies at 40° and 60°S having opposite signs, representing north–south fluctuations in the position of the zonal-mean midlatitude jet about its time mean position at 50°S (Kidson 1988; Karoly 1990; Hartmann and Lo 1998). This leading mode of the midlatitude zonal-flow variability observed in the SH is also readily simulated in numerical models (Robinson 1994; Yu and Hartmann 1993; Robinson 1996; Limpasuvan and Hartmann 1999). This jet shift mode is also frequently referred to as the zonal index variations (Lorenz and Hartmann 2001; Gerber and Vallis 2007).

The synoptic transient eddy momentum fluxes have a positive feedback on the zonal-mean wind and this feedback between the zonal flow and eddy forcing acts to increase the persistence and low-frequency variance of the zonal index, which was suggested and confirmed by many observational studies (Shiotani 1990; Karoly 1990; Hartmann 1995; Kidson and Sinclair 1995; Feldstein and Lee 1998; Hartmann and Lo 1998). Furthermore, the importance of eddy momentum flux–zonal flow feedbacks has been studied by various modeling studies as well. Yu and Hartmann (1993) suggested that the long-term zonal wind variations result from a strong eddy momentum flux–zonal flow feedback by diagnosing outputs from a multilevel primitive equation model. Robinson (1994), using a two-level model, found that eddy momentum fluxes act as a positive feedback on very-low-frequency (periods longer than 30 days) imposed variability in the zonal index. Lorenz and Hartmann (2001) diagnosed the National Centers for Environmental Prediction–National Center for Atmospheric Research (NCEP–NCAR) reanalysis data and confirmed such a positive eddy momentum flux–zonal flow feedback on low-frequency variability; its strength and effects on zonal wind were estimated by a simple linear model as well.
These studies have suggested that the zonal index mode is maintained by this positive eddy–zonal wind feedback such that it has a relatively long time scale (see also Yang and Chang 2007). Nevertheless, it is found that this time scale is often longer in general circulation model (GCM) and idealized model experiments than in observations or in atmospheric analysis data. Hartmann and Lo (1998) found that the time scales of the variations related to the zonal index mode in the SH is about 10 days for the unfiltered data for all seasons by using European Centre for Medium-Range Weather Forecasting (ECMWF) analysis data. Feldstein (2000) used the NCEP–NCAR reanalysis to estimate the e-folding time associated with the intraseasonal zonal index time series at around 18 and 14 days for the Northern Hemisphere (NH) winter and SH summer, respectively. Lorenz and Hartmann (2001) suggested that the observed e-folding time scale of the zonal index in the SH is around 13 days. For modeling work, Gerber et al. (2008a) suggested that the zonal index time scale is systematically overestimated by phase 3 of the Coupled Model Intercomparison Project (CMIP3) models, especially in the SH summer. The situation is even worse for idealized models. Yu and Hartmann (1993) suggested that the zonal index in the SH varied on time scales of hundreds of days in a multilevel primitive equation model, and these long-term variations are maintained by eddy forcing. In the particular cases of Polvani and Kushner (2002) and Kushner and Polvani (2004), the decorrelation time scale of their dry model’s leading annular mode is extremely long (200–500 days). Gerber and Vallis (2007) and Gerber et al. (2008b) also showed that the annular-mode time scale was several times larger than those observed in dry models driven with the Held–Suarez (Held and Suarez 1994) forcings. Thus, time scales of zonal index calculated from various models are much longer than those found in reanalysis or observations. In our idealized model simulations (to be discussed below), we also find that the persistence of zonal index is quite long and is much longer than those found in GCM runs. What is the reason for this substantial difference in the time scale?

Our hypothesis is that one possible contributing factor is the lack of moist processes in these idealized dry model experiments. As we mentioned in the paragraphs above, many studies have shown that transient eddy momentum flux provides a positive feedback on the low-frequency zonal index variations. However, as far as we know, no study has investigated the role that diabatic heating related with eddies play on midlatitude jet variability. In this study, we will address this issue by diagnostic studies based on observation and reanalysis data, as well as conducting experiments using a dry idealized model.

In this paper, we will first provide a brief discussion of the data, the idealized model that we used, and time scale analysis (section 2). In section 3, we will present results relating the zonal index to precipitation variations based on regression studies. In section 4, results from three different sets of experiments designed to mimic the diabatic heating associated with eddies in the model are described, and the time scales of zonal index in these experiments are compared to those computed based on the dry model runs. We will then discuss the dynamics involved and possible implications in section 5. The conclusion will be presented in section 6.

2. Data and model

a. Data

For this study, we used the NCEP–NCAR reanalysis (Kalnay et al. 1996) and 40-yr ECMWF Re-Analysis (ERA-40) (Uppala et al. 2005) four-times-daily wind and temperature data on constant pressure levels. We used data for the SH summer [December–February (DJF)] from December 1979 to February 2002 for ERA-40, and to February 2009 for NCEP–NCAR reanalysis, on a 2.5° × 2.5° latitude–longitude grid and 17 vertical levels (1000, 925, 850, 700, 600, 500, 400, 300, 250, 200, 150, 100, 70, 50, 30, 20, and 10 hPa). We also used NCEP–NCAR and ERA-40 precipitation rates to estimate the latent heating.

With respect to precipitation data, the Global Precipitation Climatology Project (GPCP) 1° daily (December 1996–February 2009, Huffman et al. 2001) and monthly (December 1979–February 2001, Huffman et al. 1997; Adler et al. 2003) precipitation are used as observations.

The wind, temperature, and precipitation of a 100-yr seasonal climatological run of the Community Atmospheric Model version 3.1 (CAM3.1), as well as model results from 17 CMIP3 models from 12 modeling centers (daily-mean data from 1961 to 2000) were also examined in this study. The list of CMIP3 models examined can be found in appendix A.

b. Idealized dry model

In addition to analyzing all of the data sources described above, we used an idealized dry model (Chang 2006) to conduct some idealized experiments. This model is based on the dynamical core of the Geophysical Fluid Dynamics Laboratory (GFDL) global spectral model (Held and Suarez 1994). The model is run with horizontal resolution of T42 (~2.8° grid spacing) and 20 evenly distributed sigma levels in the vertical. Realistic orography smoothed to model resolution is imposed. A
land–sea mask is used with stronger surface friction over land. The only other physical processes represented in the model is Newtonian damping to a radiative equilibrium potential temperature profile \( \theta_E \), together with scale selective diffusion. The thermodynamic equation can be written as

\[
\frac{D\theta}{Dt} = -\frac{\theta - \theta_E}{\tau} - \kappa \nabla^8 \theta. \tag{1}
\]

To enhance the amplitude of eddies to be close to that of the atmosphere, the target climate is imposed with the observed temperature profile (here we used the NCEP–NCAR reanalysis January climatology) but with reduced static stability to mimic the impact of moist effects in amplifying eddy amplitudes in a dry model, and the heating \( \dot{Q} \) (which is fixed for each model run) is iteratively tuned until a model climate close to \( \theta_C \) is achieved, as follows:

\[
\theta_E = \theta_C + \tau \dot{Q}, \tag{2}
\]

\[
\theta_C = \theta_{\text{obs}}(x, y, p) - A z(p), \tag{3}
\]

\[
Q_N = Q_N^{-1} - \frac{2}{3} \frac{\theta_N^{-1} - \theta_C}{\tau}, \quad N = 1, 2, 3, \ldots. \tag{4}
\]

More details concerning the model formulation can be found in Chang (2006). In this study, the control run is made with \( A = 0.65 \text{ K km}^{-1} \) in Eq. (3). Chang (2006, 2009) has shown that this model has good climatology of both the mean flow and the storm tracks (i.e., eddy statistics) when forced with the appropriate heating and that the heating \( \dot{Q} \) required to achieve this resembles the heating derived from reanalysis data based on the residual computed from the thermodynamics equation (e.g., Held et al. 2002).

c. Decorrelation time scale

To address the problems of the very long time scale of low-frequency variations occurring in our idealized model simulations, we performed the time-scale analysis of the zonal index mode in the SH based on reanalysis data, observations, GCM, and idealized dry model experiment results. Gerber et al. (2008b) presented a simple procedure to compute the \( e \)-folding time scale of the annular-mode autocorrelation function, which can concisely quantify the persistence of the low-frequency variability in a model experiment and be computed easily in practice. In this procedure, an empirical estimate \( \tau_N \) of the “true” autocorrelation time scale for \( N \) days is computed by fitting the autocorrelation function \( r_N(t) \) to a simple exponential. In particular, \( \tau_N \) is chosen so as to minimize the rms distance between \( r_N(t) \) and \( \exp(-t/\tau_N) \) for all \( t \) such that \( r_N(t) > e^{-1} \). Using this methodology, we calculated the autocorrelation time scale of the zonal index mode (also known as jet shift mode), as well as that of the eddy momentum flux and precipitation pattern related to the SH jet shift. In most cases, the empirical autocorrelation computed from each dataset is also shown.

In this study, the reanalysis data used ranges from December 1979 to February 2002 (DJF), consisting of 2076 days. The GPCP daily precipitation only has daily data from 1996 to 2009 for the NH winter (DJF), giving a total of 1173 days. Given that these datasets do not cover a long time range, the long-time-scale (>15 days) correlation may not be reliable (see Simpson et al. 2010).

To reduce such uncertainty, we have used much longer time-period data from our idealized dry model experiments (about 8500–18 000 days) and CMIP3 multimodel runs (1961–2000 DJF consisting of around 3600 days).

3. Relationship between jet shift and precipitation

Guo (2010) found that the zonal-mean precipitation is significantly correlated with the zonal index (jet shift) mode. We also find similar precipitation patterns in reanalysis data (both NCEP–NCAR and ERA-40), satellite observations (GPCP), and GCM simulations (CAM 100-yr run and CMIP3 multimodel runs) as well.

Figure 1 shows the regression of daily zonal wind \( u \) at 300hPa \( (u_{300}) \) and eddy momentum flux \( (\dot{u}v; \dot{u}v_{300}) \) against the principal component (PC) of 300-hPa zonal wind first EOF mode in the SH based on several data sources including NCEP–NCAR reanalysis, ERA-40, CAM 100-yr-run results, and the control run of our idealized model. From this figure, it is clear that the zonal wind anomalies display a poleward shift pattern (recall that the climatological position of the jet is around 50°S), with the opposite phase corresponding to an equatorward shift. The eddy momentum flux anomaly also acts consistently in terms of convergence at the poleward side of the jet and divergence at the equatorward side of the jet. It is of interest to note that, while the EOF patterns based on the two reanalysis datasets are quite similar (Fig. 1a), the momentum flux corresponding to the EOF is weaker in the NCEP–NCAR reanalysis (Fig. 1b). This is because SH transient eddy activity in the NCEP–NCAR reanalysis is biased weak owing to the impacts of satellite retrieved temperature data (see Guo et al. 2009).

In our control experiment, the EOF1 pattern displays a secondary peak in the subtropics, and correspondingly the eddy momentum flux anomaly also displays a second peak corresponding to equatorward flux from the mid-latitudes. Consistent with previous studies (e.g., Lorenz

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and Hartmann 2001; Yang and Chang 2007), EOF1 accounts for about 37% of the variance in the SH in reanalysis data. In our control experiment, EOF1 accounts for 41% of the variance, and it accounts for 43% of the variance in the CAM experiment.

To examine whether a robust relationship exists between the precipitation and jet variability, we regress the PC of the first ERA-40 EOF mode of the SH zonal-mean zonal wind at 300 hPa (zonal index) on the daily precipitation anomalies of the GPCP satellite observations, ERA-40, and NCEP–NCAR reanalysis. This precipitation pattern comparison is shown in Fig. 2a. The amplitude based on the two reanalysis datasets is quite similar with each other, while the reanalyses have a much larger magnitude than the satellite observations (GPCP) at higher latitudes. However, note that GPCP data tend to be less reliable at high latitudes (see Adler et al. 2012). In any case, the pattern is very consistent among these data sources: enhanced precipitation centered near 60° and 25°S and reduced precipitation centered near 45°S. This agreement among three datasets indicates that there exists a robust relationship between the jet shift and precipitation anomalies in the observations and in the reanalysis datasets. Similar patterns have also been found in most CMIP3 model experiments (not shown).

The time series of eddy momentum flux, which is calculated by projecting the pattern shown in Fig. 1b onto the original 6-hourly eddy momentum flux anomaly fields for both NCEP–NCAR reanalysis and ERA-40, are regressed against the precipitation anomalies from the same dataset to get the regression of precipitation associated with eddy momentum flux shown in Fig. 2b. The precipitation anomaly associated with the momentum flux anomaly is very similar to that shown in Fig. 2a. In both figures, the precipitation increases on the

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**Fig. 1.** Regression based on the PC of EOF1 of SH daily $u$ at 300 hPa of (a) zonal-mean daily zonal wind at 300 hPa ($u_{300}$, m s$^{-1}$) and (b) zonal-mean northward momentum flux at 300 hPa ($u'w'_{300}$, m$^2$ s$^{-2}$). Results from NCEP–NCAR reanalysis are shown in blue, ERA-40 are shown in red, control experiment of dry model are shown in black, and CAM3 are shown in green. The black dashed lines show regressions computed from the control experiment based on the time series generated by projecting the ERA-40 $u_{300}$ EOF1 pattern onto the $u_{300}$ anomalies found in the control experiment.

**Fig. 2.** Regression of daily precipitation (mm day$^{-1}$) for (a) NCEP–NCAR reanalysis based on PC of EOF1 of SH daily $u$ at 300 hPa (blue), ERA-40 based on PC of EOF1 (red), and GPCP based on ERA-40 PC1 (purple dashed); (b) NCEP–NCAR reanalysis (blue) and ERA-40 (red) based on time series computed using zonal-mean eddy momentum flux pattern shown in Fig. 1b.
poleward side of the midlatitude jet and decreases on the equatorward side, indicating more latent heat released at the poleward side of jet and less at the other side. Thus, the effect of latent heat due to the precipitation pattern shown in Fig. 2 is similar to the effect of eddy heat flux: transporting heat from the warm side to the cold side of the jet, making the temperature gradient weaker. Therefore, the latent heat release of precipitation related with eddy anomalies acts to reduce the baroclinicity and tends to decrease the eddy generation, opposing the impact of the eddy momentum forcing. Further discussions of this feedback will be given in section 5. Figure 2 shows that similar precipitation anomalies are related to both the jet shift and the eddy momentum flux anomalies. The question is whether the precipitation pattern is more directly related to the midlatitude jet shift or the eddy momentum flux anomalies.

To explore this, we calculate three time series by projecting the regressions of $u_{300}$, $w_{300}$, and precipitation anomalies, all based on the PC1 of SH $u_{300}$ (shown in Figs. 1a, 1b, and 2a, respectively), onto their original fields taken from the NCEP–NCAR reanalysis for the winters of 1979–2001. The autocorrelation of these three time series then is computed to examine the time scales of the different anomalies. The same calculation is done for GPCP observations for 1996–2008. The autocorrelation of 300-hPa zonal wind, eddy momentum flux, and precipitation based on the SH zonal index are plotted for both NCEP–NCAR reanalysis and GPCP observations (Fig. 3). It is clear that the time scale of precipitation (less than 5 days) is much closer to that of the eddy momentum flux (less than 5 days) than that of the zonal wind (about 12–14 days) in both reanalysis data and the GPCP observations. However, the lag correlations between the GPCP precipitation time series and both zonal wind and eddy momentum flux time series peak at a value of around 0.2 (significant at >99% based on a Student’s $t$ test); hence, in our experiments below, we will use both the zonal wind and eddy momentum flux anomalies to specify the latent heating anomaly.

4. Experiments with extra diabatic heating

Previous research suggested that the time scale of zonal index variations in GCM simulations is somewhat longer than that in reanalysis data (Gerber et al. 2008a), and even longer time scales are found in our idealized dry model simulations. In Fig. 4, the autocorrelation of the PC1 time series of the SH zonal index for NCEP–NCAR reanalysis, ensemble mean of CMIP3 climate model experiments, and our idealized dry model control experiment have been compared. The NCEP–NCAR reanalysis can be considered to represent the real atmosphere, the GCM simulation shows a longer time scale (about 30 days) and our model’s control run has an even longer time scale (about 60 days). Although the reasons for the time scale bias for GCMs are still not well understood, we hypothesize that the difference between the GCMs and our control run may be partly due to the lack of moist effects in our dry model, and we examine this in our modeling studies.

Previous studies have shown that significant positive feedback via changes in the eddy momentum flux acts to prolong the $\epsilon$-folding time scale of the zonal index, while eddy sensible heat flux acts to damp this mode (Yang and Chang 2007). In this study, we will explore the impacts of diabatic heating associated with the eddies on
the time scale of the zonal index. To do that, we will add extra diabatic heating forcing onto our dry model experiments to mimic some of the effects of moisture. In section 3, we found a robust relationship between the precipitation and eddy/jet anomalies associated with the zonal index in reanalysis data, observations, as well as in GCM simulations. The extra heating derived from such a precipitation distribution will be imposed in our idealized dry model to study the impact of diabatic heating.

a. Constant heating

The first set of experiments that we conducted is to impose a constant extra forcing at each time step in our idealized model. This constant forcing is a representation of the diabatic forcing due to latent heat release of the precipitation distribution we found in the previous section (Fig. 2b). The idealized heating is based on the precipitation anomaly found in the NCEP–NCAR regression, and is shown in Fig. 5a. The zonally symmetric extra diabatic forcing is imposed in the SH and consists of warming in the region of increased precipitation and cooling in the region of decreased rain due to eddy anomalies associated with the zonal index. In the vertical direction, the extra heating is equally distributed in the layer from 850 to 350 hPa in which the large scale condensation processes mainly occur. The vertically integrated heating corresponds to the precipitation rate shown in Fig. 2b.

After imposing the constant extra latent heat due to precipitation associated with the SH zonal index related eddy momentum flux at 300 hPa similar to the real atmosphere, the midlatitude jet moves equatorward. As shown in Fig. 5b, the green line is the zonal wind difference at 300 hPa between the constant-forcing experiment and the control run, and the black line is the zonal wind regression of the control run based on its SH zonal index (PC of EOF1). The zonal wind difference for all levels is shown in Fig. 5c, with anomalies peaking around the layer from 200 to 300 hPa. Throughout the model atmosphere, the zonal wind accelerates on the equatorward side of the jet and decelerates on the poleward side of the jet, suggesting that the midlatitude jet is forced to shift equatorward by this heating.

Given that the latent heat of precipitation we imposed in this study is associated with eddy momentum flux anomalies associated with poleward shift of the jet and after imposing the extra heating the jet moves equatorward, this diabatic forcing clearly acts to oppose the effects of eddy momentum forcing and damps the zonal wind anomaly. As mentioned before, the eddy momentum flux has a positive feedback on midlatitude jet variations (Lorenz and Hartmann 2001), therefore in this set of experiments we see that the diabatic heating related to eddy anomalies gives rise to a negative feedback on the jet shift mode.

To test the sensitivity of the response to the vertical structure of the heating, we have also imposed heating with a vertical profile similar to the heating function shown in Fig. B1b, which has a peak in the lower troposphere and decreasing linearly to zero at the 300- and 900-hPa levels, but with the same vertically integrated
total heating. The results (not shown) are nearly identical to those shown in Fig. 5. We have also examined the response to heating having the opposite sign to that shown in Fig. 5a, and, as expected, the jet responds by shifting poleward (not shown).

b. Time-varying heating

The constant forcing discussed in the preceding subsection is fixed and time independent. However, in the real atmosphere the diabatic forcing cannot be constant; it must fluctuate in time following the precipitation anomalies related to the eddy and jet anomalies. Therefore, in our second set of experiments, the extra heating imposed is time dependent based on the idealized dry model zonal-mean zonal wind or momentum flux anomalies at each previous time step.

In the first set of experiments, at each time step, zonally symmetric heating is imposed based on the pattern shown in Fig. 5a. However, the amplitude of the heating is not constant but varies depending on the jet anomaly at the previous time step. Specifically, the regressed pattern of zonal-mean $u_{300}$ related to the zonal index variation based on ERA-40 is used as a reference. This pattern (see Fig. 1a) is projected onto the model’s zonal-mean $u_{300}$ anomaly at the previous time step to obtain an amplitude (that can be either positive or negative). This amplitude is then used to set the amplitude of the diabatic heating anomaly at the current time step. So the derived latent heating associated with such a time-varying precipitation pattern imposed in the model is mimicked in the model.

The leading EOF for this experiment is shown in Fig. 6a (cyan line) and the regression of $u_{300}$ with PC1 in Fig. 6b. EOF1 now accounts only for 25% of the variance, and still somewhat resembles a jet shift, but the negative part of the dipole now extends all the way into the tropics. When the amplitude of the heating is increased by a factor of 2 (not shown), the leading EOF no longer resembles a jet shift, with the pattern dominated by a peak in the tropics. In this case, the third-leading EOF is the one that resembles the jet shift best and it only accounts for 17% of the total variance.

Since the zonal index is no longer the first-leading mode in some of the experiments after imposing extra heating, here we project the first EOF pattern of ERA-40 $u_{300}$ (shown in Fig. 1a, which, as discussed above, is the zonal index mode) onto the model-simulated zonal-mean zonal wind anomalies at 300 hPa to get the time series from which the autocorrelation of the SH zonal index for the model runs is calculated. Note that our results are not sensitive to this projection: autocorrelation time scales computed based on the EOF that best resembles the jet shift mode are very similar to these results (not shown). We decided to use this projection to generate time series for all experiments for comparison because the structure of the heating that we impose is based on jet shift derived from reanalysis data. As an example, the regression of $u_{300}$ based on this regressed time series for the control experiment is shown as the black dashed line in Fig. 1a. This pattern now resembles the patterns derived from the reanalysis data slightly better, with the amplitude of the secondary peak in the subtropics significantly reduced.

The model’s autocorrelation of the zonal index after imposing extra latent heating is plotted in Fig. 7a (solid cyan line), and the black line shows the autocorrelation based on the control run without extra heating. The dashed cyan line corresponds to the case in which the amplitude of the extra heating is reduced by 50%. Figure 7a shows that the time scales for the extra heating runs are
shorter than that of the control run, and they get shorter when larger amplitudes of extra heating are applied. The time scale becomes even shorter for the experiment in which the amplitude of the heating is doubled (not shown).

A second set of experiments has been conducted, with the amplitude of the heating imposed at each time step based on the projection of the model’s zonal-mean $u_{300}$ anomaly onto the $u_{300}$ pattern shown in Fig. 1b at the preceding time step. The leading EOF for this experiment, which accounts for 37% of the total variance, is shown by the orange line in Fig. 6, while the autocorrelation for the zonal index time series derived from this experiment is shown by the orange line in Fig. 7a. We can see that imposing heating based on momentum flux anomalies also acts to reduce the zonal index time scale.

c. Mimicking condensational heating in regions with upward motion

Other than the previous experiments of imposing diabatic heating derived from precipitation associated with jet and eddies anomalies, another way to mimic the diabatic heating in a dry model is to parameterize the condensational heating based on model upward motion (e.g., Becker and Schmitz 2001). In this study, the self-induced condensational heating in the extratropics is scaled with the pressure velocity $\omega$ and is only active in regions of rising motion, indicated by the Heaviside step function $H(-\omega)$, as follows:

$$\frac{D\theta}{Dt} = -\frac{\theta - \theta_E}{\tau} + C(x, y, p) \omega |H(-\omega)|.$$  (5)

The strength of the heating is derived from regression between daily condensational heating and $\omega$ from a seasonal full physics GCM run made with the GFDL R30 spectral model with 14 vertical levels (Alexander and Scott 1996; Gordon and Stern 1982). More details about the model, and the spatial structure of the function $C(x, y, p)$, are shown in appendix B. Since in Eq. (5) the impact of diabatic heating in enhancing baroclinic waves has already been included in the newly added term, in this experiment there is no need to reduce the static stability of the model atmosphere, so the parameter $A$ in Eq. (3) is set to zero. The heating $Q$ is again iterated with this extra time-varying heating added until the model climate becomes close to the imposed $\theta_C$ (with $A = 0$).

The leading EOF for this experiment is shown as the purple dotted line in Fig. 6, and it accounts for 32% of the total variance. In Fig. 7b, the autocorrelation for two different runs has been compared in the same way as in the previous subsection. The gray line is the autocorrelation of zonal index variation of a control run without reduction of static stability [$A = 0$ in (3)], while the dotted purple line is the run forced to the same basic
state but with parameterized “condensational heating” given by Eq. (5). It is clear that the time scale of SH zonal index is substantially reduced by adding the condensational heating in the upward-motion regions, compared with that of the control run without any reduction of static stability. For the model control run, the autocorrelation curve is quite flat with an $e$-folding time much longer than 60 days, suggesting that a very-low-frequency variation dominates the jet shift. However after adding the condensational heating in the regions with rising motion, the $e$-folding time becomes much shorter and is about 17 days—much closer to the time scale derived from reanalysis data. This result suggests that the jet shift mode is significantly damped with parameterized condensational heating in regions of rising motion. It is of interest to note that the time scales of the second EOF mode (jet strengthening or broadening) and the third EOF mode (related to the subtropical jet) are also substantially reduced in this experiment (not shown).

5. Discussion

a. Feedback mechanisms

Here we briefly summarize the differences in the feedbacks related to eddy momentum flux, eddy heat flux, and diabatic heating. When the midlatitude jet shifts poleward, this results in the poleward shift of the baroclinic zone. The wave source of synoptic eddies shifts with the jet. More waves then propagate equatorward from the wave source at higher latitudes, thus giving rise to more momentum fluxes converging into the new jet position to reinforce the wind anomalies (Lorenz and Hartmann 2001). The forcing anomalies also drive a meridional circulation that reinforces the low-level baroclinicity (Robinson 2000), thus enhancing the eddy source over the anomalies. This is the positive feedback of the eddy momentum flux on zonal wind anomalies. On the other hand, the eddy heat flux transports heat poleward around the wave source (upward EP flux is equivalent to poleward eddy heat flux), and it mixes the air around the cyclones by warming on the poleward side and cooling on the equatorward side to weaken the temperature gradient (Yang and Chang 2007). In this way, it weakens the wave source and works against the effect of the eddy momentum forcing. Therefore, the heat flux has a negative feedback on the zonal wind anomalies. Regarding the effect of diabatic heating investigated in this study, we find the latent heat of condensation related to precipitation associated with eddy anomalies due to jet shift has a similar effect as the eddy heat flux. Based on Fig. 2, the positive precipitation anomalies due to jet shift are poleward of the midlatitude jet core in the SH. Thus, there is warming at the poleward side of the jet and cooling at the equatorward side, forcing a weaker temperature gradient to weaken the baroclinic wave source, and thus provides a negative feedback to the impact of the eddy momentum flux.

In section 4b, we imposed extra heating based on jet and eddies anomalies associated with the zonal index in our dry model and find that this forcing works against the effect of eddy momentum forcing to give rise to a negative feedback on jet variations and shortens the zonal index time scale. This is the direct impact of diabatic heating on jet variations. Apart from this direct effect, the experiments discussed in the following paragraphs suggest that there may be another indirect mechanism behind the relationship between diabatic heating and the zonal index time scale.

In Fig. 7c, results from five different experiments using different values of $A$ [see Eq. (3)] ranging from 0 to 2.0 K km$^{-1}$ are shown. For each experiment, the heating ($Q$) is recomputed by iteration such that the climatological temperature structure (apart from the static stability) is similar to that observed. In these experiments, no diabatic feedback to the jet variation is imposed. The results show that, as the static stability is reduced, the time scale is shortened. Why is that the case? Our hypothesis is that even though in these experiments, no diabatic feedback is present, as the static stability decreases, baroclinicity increases, resulting in enhanced eddies. The enhanced eddies then make the system more chaotic, and in this way the zonal index time scale is reduced.

In the experiment in which we add the condensational heating in the upward motion regions (section 4c), since rising motion generally occurs over the warm sector of cyclones, the warm sector of eddies tends to get warmer, resulting in increased eddy available potential energy. Thus, the eddy amplitude is enhanced (Gutowski et al. 1992; Emanuel et al. 1987) and the system becomes more unstable and chaotic, and the time scale is further reduced in the same way as the zonal index time scale is reduced in the reduced static stability experiments. This effect on the zonal index time scale can be regarded as an indirect effect of diabatic heating. This can potentially explain why the parameterized heating experiment discussed in section 4c appears to be more effective in reducing the zonal index time scale than the experiments discussed in section 4b. In the former case, both direct and indirect effects are present, while in the latter case, only the direct effect is present while the indirect effect is not, since in these experiments the imposed extra heating is zonally symmetric and thus is not effective in enhancing eddy amplitude. On the other hand,
for the experiments with reduced static stability but without the imposed diabatic feedback, only the indirect effect (enhanced eddy amplitudes) is present while the direct effect is not, hence in these experiments the time scales are also not reduced as much as that for the condensational heating experiment (cf. Fig. 7c to 7b).

The separate impacts of eddy amplitude and diabatic heating discussed in the preceding paragraph can be illustrated by Fig. 8a. In this figure, the zonal index time scale for each experiment shown in Fig. 7 is plotted against the eddy amplitude. Eddy amplitude is indicated by the hemispheric-mean (20°−70°S) 300-hPa eddy meridional velocity variance computed using a 24-h difference filter (Wallace et al. 1988) that highlights synoptic-time-scale transient eddies with periods between 1.2 and 6 days.

In Fig. 8a, results from the five experiments with different static stability discussed above (Fig. 7c) are shown by the black asterisks. For these five experiments, as the static stability is decreased, eddy amplitude increases, and the zonal index time scale decreases.

Results from the experiments with extra heating based on jet shift anomalies (section 4b, Fig. 7a) are shown by the filled cyan triangles. For these experiments, since the amplitude of the heating varies with the jet anomalies and takes on both positive and negative values over time, these heating anomalies largely cancel out and, thus, in these experiments the mean static stability and eddy amplitude are both not significantly changed. These results should be compared to that of the control experiment indicated by the black unfilled triangle. Comparing these three triangles, we can see that the extra heating acts to reduce the zonal index time scale. Similarly, results for the experiment with additional heating based on eddy momentum flux anomalies (orange triangle) also show the impact of this extra heating in reducing the zonal index time scale.

Results from the experiment in which heating is added in the regions with upward motion (section 4c, Fig. 7b) are shown by the filled purple circle. Since in this experiment, \( A \) in (3) is set to zero, these results should be compared to those with the same static stability but without the added heating (black unfilled circle). It is clear from Fig. 8a that, in this experiment, the heating acts to greatly increase the eddy amplitude and hence decrease the zonal index time scale. However, comparing with the results of the five experiments with no diabatic heating feedback (black asterisks), the diabatic feedback also acts to further decrease the zonal index time scale from the value expected based on the eddy amplitude alone. Since the results indicated by the purple and black circles are experiments forced to the same static stability and they exhibit very different zonal

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**FIG. 8.** (a) Decorrelation time of zonal index (days) plotted against climatological SH-mean 300-hPa eddy meridional velocity variance statistics (m²/s²) for all experiments shown in Fig. 7. The asterisks show results from the experiments with different values of \( A \), the triangles show results from experiments with heating based on jet shift anomalies (cyan) and momentum flux anomalies (orange), and the circles show results from experiments with heating added in regions with rising motion. The red square symbol shows results from ERA-40 data. (b) As in (a), but for CMIP3 models (blue) and ERA-40 (red squares). (c) Average value of day-5–10 autocorrelation for PC1 of SH 300 hPa \( u \), plotted against the minimum value of zonal-mean eddy momentum flux (m²/s²) for CMIP3 models (blue symbols) and ERA-40 (red square symbol).
index time scales, we believe that the static stability itself does not play a direct controlling role in determining the zonal index time scale, but rather acts indirectly through its impact on changing the eddy amplitude.

The results discussed in the preceding paragraphs suggest that the zonal index decorrelation time scale should be inversely correlated to eddy amplitude. To test this possibility, we have computed the decorrelation time scale of the zonal index mode for the SH summer from twentieth-century simulations made by 17 CMIP3 models (see appendix A for the list of models), together with the eddy amplitude as defined above. The two quantities are plotted against each other in Fig. 8b. It is clear that these two quantities are negatively correlated, with a correlation of \(-0.60\), which is highly significant (\(-99\%\) significant based on a two-tailed Student’s \(t\) test)—this correlation is still significant at the 95% level even if models from the same modeling center are treated as not independent). Given that some of the time scales shown in Fig. 8b are relatively long, we have used an alternative indicator of the autocorrelation time scale by computing the geometric mean value of the daily autocorrelation from day 5 to day 10. This quantity is plotted against the maximum negative value of the eddy momentum flux in Fig. 8c. Again, Fig. 8c shows that the stronger the eddy momentum flux, the lower the value of the autocorrelation. Nevertheless, Fig. 8 also indicates that the eddy amplitude is clearly not the only factor affecting the zonal index decorrelation time scale in CMIP3 simulations, as all model simulations, including those that have eddy variance and/or momentum flux that are stronger than that found in ERA-40, have longer decorrelation times than those found in ERA-40 data, consistent with the results of Gerber et al. (2008a) that the zonal index time scale for SH summer is systematically overestimated by CMIP3 models.

b. Possible implications

As discussed above, diabatic heating associated with the eddies tends to damp the zonal index variations and shorten the zonal index time scale. The importance of correctly simulating the zonal index time scale has been demonstrated by Gerber et al. (2008b), who showed that the amplitude of their idealized model’s response to external forcing increases as the time scale increases, as suggested by the fluctuation–dissipation theorem (Leith 1975). Here, we have conducted a series of experiments to examine whether similar results also hold for our model under a different external forcing than that imposed by Gerber et al.

Gerber et al. (2008b) conducted their perturbation experiments by imposing an anomalous momentum forcing following Ring and Plumb (2007). Here, we will impose diabatic forcing similar to those imposed by Butler et al. (2010) to qualitatively mimic aspects of climate change due to increased greenhouse gas forcing, including warming in the tropical troposphere, cooling in the polar stratosphere, and warming at the polar surface. Here, instead of using idealized analytical forcings like Butler et al. did, we have chosen to apply the forcing by imposing a \(\Delta \theta\) on \(\theta_e\) of Eq. (3) equal to the mean temperature change projected by an ensemble of CMIP3 models (models included in the mean are specified in appendix A). The imposed \(\Delta \theta\) is three dimensional, and the zonal mean (in terms of \(\Delta T\) instead of \(\Delta \theta\)) is shown in Fig. 9a. Note that the radiative damping time scale \(\tau\) used in Eq. (1) is 30 days in the free atmosphere; hence the forcing has a maximum amplitude of about 0.2 K day\(^{-1}\) in the tropical upper troposphere, weaker than that applied by Butler et al. (0.5 K day\(^{-1}\)). The zonal-mean zonal wind change projected by the same CMIP3 model ensemble is shown in Fig. 9b for reference.

The response when the forcing is applied in our control model (\(A = 0.65\text{K km}^{-1}\) and no extra diabatic feedback) is shown in Figs. 9c,d. The zonal-mean temperature change (Fig. 9c) is qualitatively similar to the imposed change in radiative equilibrium temperature profile (Fig. 9a), and the model zonal wind response (Fig. 9d) shows a clear poleward shift of the midlatitude jets. Note that the model jet response is much stronger than that found in the CMIP3 model experiments. In Fig. 9e, the zonal wind response when the forcing is applied to the model with added time-varying heating that depends on the eddy momentum flux anomalies at the previous time step (section 4b) is shown. With the extra diabatic forcing that acts to shorten the zonal index time scale, the jet response is slightly weaker than that for the control experiment. Finally, in Fig. 9f, the response when the forcing is applied to the model with parameterized heating in midlatitude regions of rising motion (section 4c) is shown. Consistent with the results of Gerber et al. (2008b), the jet response for this model (which has a significantly shorter zonal index time scale than the control experiment) is significantly weaker and is much closer to the jet response found in the CMIP3 GCM experiments (Fig. 9b). These results suggest that jet shift response found in forced idealized dry model experiments (e.g., Butler et al. 2010; Lorenz and DeWeaver 2007; Haigh et al. 2005) must be interpreted with care since most dry models have zonal index time scales that are significantly biased long (partly because of the absence of the diabatic feedback discussed in this study) and, thus, are likely overly sensitive to climate forcing.

Note that, as in Gerber et al. (2008b), the agreement between our results and the predictions of the fluctuation–dissipation theorem is qualitative rather than quantitatively...
accurate. For the control experiment, the zonal index time scale is about 61 days, and the imposed forcing results in a jet shift of about 2.5°. For the experiment with heating added based on eddy momentum flux anomalies, the zonal index time scale is about 37 days, but the jet shift in this experiment is only slightly smaller (2.2°). For the experiment with heating added in regions of rising motion, the zonal index time scale is about 17 days (3.6 times smaller) while the jet shift under the imposed forcing is about 1° (2.5 times smaller). Thus, in our experiment, the response to forcing is not directly proportional to the zonal index time scale. Gerber et al. (2008b) discussed a number of possibilities why the fluctuation–dissipation theorem is not expected to be strictly valid in these kinds of experiments, and readers are referred to that paper for more details.

6. Conclusions

In this study, the role that diabatic heating plays in zonal index variations is investigated using diagnostic and modeling studies. First, by diagnosing satellite precipitation observations, reanalysis data, and GCM data, we have found that the precipitation distribution related to the zonal index (basically north–south shift of the jet) in the SH has a very
similar pattern among all datasets. The precipitation associated with poleward shifted (equatorward shifted) midlatitude jet increases (decreases) on the poleward side of the jet and decreases (increases) on the equatorward side. Nevertheless, the amplitude of the precipitation anomalies derived from satellite observations is weaker than that based on the two reanalysis datasets.

Since precipitation in the midlatitudes is due to baroclinic waves, we expect that the precipitation anomalies should also be related to anomalies in the eddies. Hence we computed the regression of precipitation based on eddy momentum flux at 300 hPa. The results show a similar pattern of precipitation anomalies as that derived based on the jet shift. Therefore, we concluded that the precipitation anomalies are related to anomalies in eddy momentum fluxes as well as those in the jet.

Meanwhile, we find that the zonal index variability in our dry model experiments displays a very long time scale—much longer than those found in GCM experiments. We hypothesize that one possible reason for this difference between GCM and our model results may be the lack of moist processes in the dry model. This hypothesis is tested by imposing extra diabatic forcing in our dry model in three different ways to mimic the effect of diabatic heating related to the eddy anomalies.

First, a time-independent diabatic heating is imposed in our idealized dry model. This extra heating is based on the precipitation anomaly regressed from the momentum flux anomaly associated with the poleward shift of the jet. When this heating is imposed, the midlatitude jet moves equatorward (poleward) when the precipitation is related to poleward (equatorward) shifted eddies. The results suggest that the diabatic forcing gives rise to a negative feedback on midlatitude jet variations, working against the effects of the eddy momentum fluxes.

Next, a set of experiments in which time varying diabatic forcing is added to the model is performed with the amplitude and sign of the heating imposed based on the modeled jet anomalies at the previous time step. In this set of experiments, when the amplitude of heating is large, the leading mode of 300-hPa zonal-mean zonal wind is no longer the jet shift mode. In all of these experiments with added heating, the time scale of the jet shift mode is reduced, and the percentage of variance the jet shift mode explains decreased, suggesting that this north–south jet shift mode is suppressed by adding the extra latent heat of precipitation related with the jet shift. Results of another set of experiments in which heating is added based on eddy momentum flux anomalies also give similar results. Given these results, we can say that the zonal index is damped by a negative feedback of the diabatic forcing. The time scale of jet shift mode in these experiments with extra heating is shorter than that of the control run and it generally gets shorter when the magnitude of the extra heating is enhanced.

The third way to mimic the effect of diabatic heating in the dry model is to parameterize diabatic heating as a function of $-\omega$ in regions where there is upward motion. In this experiment, the jet shift mode was significantly damped. The time scale of the SH zonal index becomes much closer to that observed.

Apart from the direct impact diabatic heating has on damping the jet shift, our experiments suggest that diabatic heating has another indirect impact. In a series of experiments in which eddy amplitudes are systematically changed by changing the static stability of the radiative equilibrium temperature profile, the results show that, as the static stability decreases, the eddies become stronger and more chaotic and the zonal index time scale decreases. This inverse relationship between eddy amplitude and zonal index time scale is also found in the CMIP3 ensemble that we have examined. In midlatitude eddies, diabatic heating generally occurs in rising air over the warm sector of the eddies, enhancing the eddy available potential energy, making the eddies stronger and the system more chaotic. Hence, our results suggest that diabatic heating can affect zonal index variations also through invigorating the eddies.

In conclusion, our results suggest that diabatic heating provides a negative feedback to zonal index variations, partially offsetting the positive effect of eddy momentum flux, leading to a shortening of the zonal index time scale. This feedback is entirely missing in dry model experiments, partly explaining why the zonal index time scale in these experiments is much longer than that observed. Corresponding to the long time scales, our results also suggest that these dry mechanistic models may be overly sensitive to climate forcing that produces a jet shift response.

With atmospheric moisture projected to increase under global warming, it would be of interest to investigate whether this diabatic feedback will be enhanced and impact the zonal index time scale in the future.

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**APPENDIX A**

**List of CMIP3 Models Examined**

In this study, daily data from the twentieth-century experiment from 17 CMIP3 models have been examined. The models are CGCM3.1(T47), CGCM3.1(T63),
CNRM-CM3, CSIRO Mk 3.0, CSIRO Mk 3.5, GFDL CM2.0, GFDL CM2.1, GISS-AOM, GISS-EH, GISS-ER, INM-CM3.0, IPSL-CM4, ECHO-G, ECHAM5/MPI-OM, and MRI-CGCM2.3.2 [the models (and institution) are listed in Table A1]. For each model, data from a single run have been examined. Some properties (including resolution and biases in storm track amplitudes) of these model simulations can be found in Chang et al. (2013).

The projected temperature and zonal wind changes shown in Figs. 9a,b are computed based on the Special Report on Emissions Scenarios (SRES) A2 and A1B scenarios, using monthly mean data from the following models: BCCR BCM2.0, CGCM3.1(T47), CGCM3.1(T63), GISS-AOM, GISS-EH, GISS-ER, MIROC3.2(hires), FGOALS-g1.0, INM-CM3.0, IPSL-CM4, ECHO-G, ECHAM5/MPI-OM, and MRI-CGCM2.3.2 [the models (and institution) are listed in Table A1]. For each model, data from a single run have been examined. Some properties (including resolution and biases in storm track amplitudes) of these model simulations can be found in Chang et al. (2013).

A previous study by Son et al. (2008) has suggested that ozone recovery may have impacts on jet shifts in the SH, and this model choice ensures more uniformity in the future projection considered.

**APPENDIX B**

**The Structure of $C(x,y,p)$**

In this appendix, the prescribed spatial distribution of $C(x,y,p)$ used in Eq. (5) is described. The strength of the heating imposed in the dry model is derived based on regression between condensational heating and $\nabla^2 v$ from a GCM run made with the full physics GFDL R30 spectral model. This is the same GCM experiment that was analyzed by Chang (2001). Details of the model formulation for a lower resolution version of the model

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**TABLE A1.** The CMIP3 models (and institutions) used in the experiments.

<table>
<thead>
<tr>
<th>Model</th>
<th>Expansion</th>
</tr>
</thead>
<tbody>
<tr>
<td>BCCR BCM2.0</td>
<td>Bjerknes Centre for Climate Research Bergen Climate Model, version 2.0</td>
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<tr>
<td>CGCM3.1(T47)</td>
<td>Canadian Centre for Climate Modelling and Analysis Coupled Global Climate Model, 3.1 (T47 version)</td>
</tr>
<tr>
<td>CGCM3.1(T63)</td>
<td>Canadian Centre for Climate Modelling and Analysis Coupled Global Climate Model, 3.1 (T60 version)</td>
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<tr>
<td>CNRM-CM3</td>
<td>Centre National de Recherches Météorologiques Coupled Global Climate Model, version 3</td>
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<tr>
<td>CSIRO Mk 3.0</td>
<td>Commonwealth Scientific and Industrial Research Organisation Mark, version 3.0</td>
</tr>
<tr>
<td>CSIRO Mk 3.5</td>
<td>Commonwealth Scientific and Industrial Research Organisation Mark, version 3.5</td>
</tr>
<tr>
<td>ECHO-G</td>
<td>ECHAM4 and the global Hamburg Ocean Primitive Equation</td>
</tr>
<tr>
<td>ECHAM5/MPI-OM</td>
<td>ECHAM5/Max Planck Institute Ocean Model</td>
</tr>
<tr>
<td>FGOALS-g1.0</td>
<td>Flexible Global Ocean–Atmosphere–Land System Model gridpoint, version 1.0</td>
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<tr>
<td>GFDL CM2.0</td>
<td>Geophysical Fluid Dynamics Laboratory Climate Model, version 2.0</td>
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<td>GFDL CM2.1</td>
<td>Geophysical Fluid Dynamics Laboratory Climate Model, version 2.1</td>
</tr>
<tr>
<td>GISS-AOM</td>
<td>Goddard Institute for Space Studies, Atmosphere–Ocean Model</td>
</tr>
<tr>
<td>GISS-EH</td>
<td>Goddard Institute for Space Studies Model E, coupled with the HYCOM ocean model</td>
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<tr>
<td>GISS-ER</td>
<td>Goddard Institute for Space Studies Model E-R</td>
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<td>L’Institut Pierre-Simon Laplace Coupled Model, version 4</td>
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<td>Model for Interdisciplinary Research on Climate, version 3.2 (high resolution)</td>
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<tr>
<td>MRI-CGCM2.3.2</td>
<td>Meteorological Research Institute Coupled Atmosphere–Ocean General Circulation Model, version 2.3.2</td>
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**FIG. B1.** The spatial distribution of the structure of $\mu g C(x,y,p)$ for $C$ given in Eq. (5): (a) horizontal structure (vertical mean) and (b) latitude–height cross section (zonal mean). Contour interval $4 \times 10^{-4} \text{K m}^{-1}$. 
can be found in Gordon and Stern (1982). Climate statistics from an integration of this model have been documented by Alexander and Scott (1996). This model experiment was used to diagnose the distribution of $C$ because we have daily three-dimensional diabatic heating values available from this run.

The structure of $C$ is qualitatively similar to the function imposed by Becker and Schmitz (2001). The horizontal distribution of the vertical average of $\rho g C$ is shown in Fig. B1a, while the vertical distribution of the zonal mean of $\rho g C$ is shown in Fig. B1b. It is clear that the heating covers regions from the subtropical jet to high latitudes and is mainly distributed in the troposphere in both hemispheres, maximizing at the position of the midlatitude jet stream and subtropical jet around 750 hPa in the SH and peaking around 35°N at the same level in the NH.

To test whether our results are sensitive to the detailed structure of $C$, we have performed sensitivity studies by broadening $C$ by two grid points ($\sim 5.6^\circ$ latitude) on both subtropical and poleward sides in both hemispheres and obtained very similar results in terms of reduction in the zonal index time scale. Therefore, we conclude that the effects of diabatic feedback in damping the jet shift mode are not very sensitive to the exact form of $C$ imposed.

REFERENCES


