Covariation of finite-amplitude wave activity and the zonal mean flow in the midlatitude troposphere: 1. Theory and application to the Southern Hemisphere summer

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Abstract Tropospheric eddy-mean flow interaction is formulated in terms of the vertically integrated budget of finite-amplitude wave activity (FAWA). At each latitude the dynamics is governed by three coupled equations for the interior and surface FAWA and barotropic zonal mean zonal flow. In midlatitude austral summer, the budget reveals a largely adiabatic, antiphase covariation of FAWA and the mean flow. A marked periodicity is found for FAWA around 20–30 days, but not the mean flow, consistent with the recently discovered baroclinic annular mode. The difference in the spectra of FAWA and the mean flow arises from (i) distinct spectra of low-level meridional eddy heat flux and the barotropic eddy momentum flux convergence and (ii) a strong thermal damping of surface wave activity: the latter makes the FAWA respond largely to the low-level meridional eddy heat flux at low frequencies, whereas the zonal mean flow responds to the momentum flux convergence whose spectrum is broader and occupies higher frequencies.

1. Introduction

Over the last two decades there has been a substantial amount of work on the variability of jet streams in the extratropical troposphere [Hartmann and Lo, 1998; Thompson et al., 2000; Lorenz and Hartmann, 2001]. The jet variability in the midlatitudes is largely eddy driven, and as such most existing work concerns the response of the mean flow to eddy forcing, but the corresponding variability in the driving eddies has received relatively scant attention.

Recently, Thompson and Woodworth [2014] and Thompson and Barnes [2014] identified a robust oscillation in eddy kinetic energy in the Southern Hemisphere extratropics with a period of 25–30 days. The oscillation is largely associated with the variability in the low-level meridional eddy heat flux and baroclinicity of the flow, rather than the eddy momentum flux, and for this reason it is termed baroclinic annular mode or BAM. (It appears that the same phenomenon had been recognized earlier by Webster and Keller [1974] with limited balloon data. Using First GARP Global Experiment (FGGE) III-b analysis, Chen et al. [1987] identified similar phenomenon and highlighted its existence in summertime.) These authors mainly concern energy conversions associated with repeated baroclinic eddy life cycles, but the energy budget requires the domain integral to close it even when the dynamics is conservative. In this article we will develop a simple diagnostic framework for eddy-mean flow interaction based on the budget of finite-amplitude wave activity (FAWA) or pseudomomentum [Nakamura and Zhu, 2010; Nakamura and Solomon, 2010, 2011]. In its barotropic form, the theory predicts that the sum of the zonal mean zonal flow and FAWA remains constant at each latitude under conservative dynamics, a direct result of Kelvin’s circulation theorem [Nakamura and Zhu, 2010]. In this limit the zonal mean zonal flow grows at the expense of the FAWA and vice versa, so their variation is antiphase. Analogous eddy-mean flow interaction in the real atmosphere entails covariation of the zonal mean zonal wind and the interior and surface FAWA at each latitude. The three quantities are driven by the interior eddy momentum flux convergence and the low-level meridional eddy heat flux but also modified by nonconservative processes such as frictional and thermal damping. This three-component model will be our baseline diagnostic theory for characterizing the eddy-mean flow interaction in the midlatitude atmosphere.

The theory is then applied to the Southern Hemisphere summer where the main driver of eddy-mean flow interaction is synoptic baroclinic eddies [Hartmann and Lo, 1998; Lorenz and Hartmann, 2001]. We will defer the discussion on seasonal and interhemispheric variabilities to a subsequent paper. We will see that a robust covariation of the FAWA and mean flow exists between 40 and 50°S at timescales around 25 days and that
the periodicity and amplitude of variation are more pronounced in FAWA than the zonal mean flow. These corroborate the previous results on the BAM. We will then show that strong thermal damping on surface FAWA, together with distinct spectral shapes of the low-level meridional eddy heat flux and the interior eddy momentum flux convergence, explains this asymmetry as well as why the wave activity variability of the BAM is dominated by the eddy heat flux. Finally, with a spectral decomposition of the FAWA budget we demonstrate that the nature of eddy-mean flow interaction varies significantly depending on the timescales. The next section outlines the diagnostic method and data sets. Section 3 describes the results, followed by a short summary in section 4.

2. Theory and Data

The FAWA is defined based on the net displacement of a conserved substance from the zonal line of equivalent latitude. In this study, we use quasi-geostrophic potential vorticity (PV) to define the interior FAWA

$$A(\phi, z, t) = \frac{1}{2\pi a \cos \phi} \left( \iint_{\phi'} q_a dS - \iint_{\phi} q_a dS \right),$$

(1)

where $a$ is the radius of the planet, $\lambda$ and $\phi$ are longitude and latitude, $z = -H \ln(p/p_0)$ is pressure pseudohigh, $dS = a^2 \cos \phi' d\phi' d\lambda$ is the area element, and $q_a(\lambda, \phi', z, t) = f + \zeta + fe^{2iH}(\theta - \tilde{\theta})/(\partial \tilde{\theta}/\partial z)\right)/\partial z$ is PV. In the calculation of PV, $f = 2\Omega \sin \phi'$ is the Coriolis parameter, $\zeta$ is relative vorticity, $\theta$ is potential temperature, $\tilde{\theta}$ is its global horizontal average, and $H$ is a constant scale height which is chosen to be 7 km. In the above, $Q(\phi, z, t)$ is chosen such that the wavy PV contour $q_a = Q$ encloses the same area as the region south of latitude $\phi$. (Note that the surface integrals extend northward from the south pole.)

Similarly, we use potential temperature to define surface FAWA

$$B(\phi, t) = \frac{1}{2\pi a \cos \phi H(\partial \tilde{\theta}/\partial z)} \left( \iint_{\phi'} \theta dS - \iint_{\phi} \theta dS \right),$$

(2)

where $\theta$ and $\partial \tilde{\theta}/\partial z$ are evaluated at the surface. When there is topography $z = \eta_0(\lambda, \phi)$, $\theta$ is replaced by $\theta_{\text{top}} + \eta_0(\partial \tilde{\theta}/\partial z)_{\text{top}}$. Note that $A \geq 0$ and $B \leq 0$ by construction. $A$ and $B$ satisfy the governing equations [Nakamura and Zhu, 2010]

$$\frac{\partial A}{\partial t} = -v^T q_a + \Lambda = \frac{1}{\cos^2 \phi \partial \phi} \left( v^T \nabla^2 \phi \right) - fe^{2iH} \frac{\partial}{\partial z} \left( e^{-2iH} \nabla^2 \phi \right) + \widetilde{A},$$

(3)

$$\frac{\partial B}{\partial t} = -f \frac{\nabla^2 \phi}{H(\partial \tilde{\theta}/\partial z)} \bigg|_{z=0} + \widetilde{B},$$

(4)

where the overbar and prime denote the zonal average and departure from it, respectively, and $\Lambda$ and $\widetilde{B}$ denote nonconservative sources and sinks of $A$ and $B$. The last expression of (3) arises from Taylor’s identity. In the quasi-geostrophic approximation the zonal mean zonal velocity obeys

$$\frac{\partial \tilde{U}}{\partial t} = f\nabla^2 \tilde{\phi} - \frac{1}{\cos^2 \phi \partial \phi} \left( v^T \nabla^2 \phi \right) + \tilde{U},$$

(5)

where $\tilde{U}$ denotes forcings such as surface friction and gravity wave drag. Defining the density-weighted vertical average

$$\langle (\cdot) \rangle = \int_0^H (\cdot) dz \int_0^H e^{-2iH} dz,$$

(6)

(3) and (5) become

$$\frac{\partial}{\partial t} \langle A \rangle = \frac{1}{\cos^2 \phi \partial \phi} \left( v^T \nabla^2 \phi \right) + \frac{f v^T \theta^T}{H(\partial \tilde{\theta}/\partial z)} \bigg|_{z=0} + \langle \Lambda \rangle,$$

(7)

$$\frac{\partial}{\partial t} \langle \tilde{U} \rangle = -\frac{1}{\cos^2 \phi \partial \phi} \left( v^T \nabla^2 \phi \right) + \langle \tilde{U} \rangle.$$

(8)
where \( \langle \bar{v} \rangle = 0 \) was assumed. From (4), (7), (8)

\[
\frac{\partial \langle \bar{u} \rangle}{\partial t} = -\frac{\partial}{\partial t} (\langle A \rangle + B) + (\langle \dot{U} \rangle + \langle \dot{A} \rangle + \ddot{B}).
\] (9)

In the absence of nonconservative effects, the barotropic zonal mean wind \( \langle \bar{u} \rangle \) increases entirely at the expense of \( \langle A \rangle + B \). In this limit, the mean flow anomaly (departure from the time-mean) and the wave activity anomaly will be opposite of each other:

\[
\Delta \langle \bar{u} \rangle = -\Delta (\langle A \rangle + B).
\] (10)

This is a finite-amplitude extension to the nonacceleration theorem, first introduced by Charney and Drazin [1961] for small-amplitude baroclinic Rossby waves.

In what follows (4), (7), and (8) are evaluated from meteorological data at each latitude and time, the nonconservative terms being the residual of the budget. We use 4 times daily reanalyses for winds and temperature from the European Centre for Medium-Range Weather Forecasts ERA-Interim data sets [Dee et al., 2011] for 1979–2013. Quasi-geostrophic PV is calculated following the procedures described in Nakamura and Solomon [2010]. The main difficulty in applying the theory to data is the treatment of the lower boundary, which is commonly raised to 850 hPa to avoid the boundary layer [Hoskins et al., 1985]. However, this invalidates (4) because the vertical advection of heat is not negligible at the top of the boundary layer, unless \( B \) is reinterpreted to absorb it. On the other hand, placing the boundary at the sea level involves (in addition to the departure from quasi-geostrophy in the boundary layer) difficulty in evaluating the surface static stability accurately, the uncertainty of which can be as large as a factor of 2. In this study we discretize Taylor’s identity in the vertical such that the boundary values are evaluated as a weighted average of the values at two lowest levels (\( z = 0 \) and 1 km). This effectively places the lower boundary within the boundary layer, with static stability more or less representing the layer-averaged value. The result is, as it turns out, not too different from
Figure 2. (top row) 250 hPa geopotential field (left) and vertical structure of zonal mean zonal wind (right) on 1800 UTC 15 December 2011. (bottom row) Same as top row but for 1800 UTC 16 January 2012. (left column) White contours are 99,000, 102,000, and 105,000 m$^2$s$^{-2}$. (right column) Contour interval is 3 m s$^{-1}$ with negative values dashed. Blue lines indicate 46.5$^\circ$S, the latitude of analysis for Figure 1.

Figure 2 delineates hemispheric atmospheric flows during a period in which wave activity is high and the mean flow is weak (top row) and another period in which the opposite is true (bottom row). These instances correspond to the two blue vertical lines in Figure 1. In the former (15 December 2011) the 250 hPa geopotential reveals a highly meandering westerly flow in the midlatitude and the core of the jet is split into polar and subtropical latitudes, with westerly winds significantly weakened around 45–55$^\circ$S.

3. Results

An example of covariation of the mean flow and wave activity is shown in Figure 1, analyzed for 46.5$^\circ$S from December 2011 to February 2012. Figure 1 (first and second panels) shows the time-height cross sections of $\Delta \bar{u}$ and $\Delta A$ (departures from seasonal mean values). While $\Delta \bar{u}$ is vertically coherent from the surface to the lower stratosphere [Thompson et al., 2000], $\Delta A$ is more localized to the upper troposphere as a result of concentrated PV gradients near the tropopause [Nakamura and Solomon, 2010, 2011]. Despite the different structures, once vertically averaged, $\Delta \bar{u}$ and $\Delta (\langle A \rangle + B)$ largely compensate each other, as shown by the black and red curves in Figure 1 (third panel). This is consistent with (10). Because of its negative values, surface wave activity $B$ is supposed to offset some of the interior wave activity $\langle A \rangle$ and its anomaly. However, the very small difference between $\Delta (\langle A \rangle + B)$ (red) and $\Delta \langle A \rangle$ (cyan) indicates that $B$ fluctuates little ($B$ itself is also much smaller than $\langle A \rangle$). More importantly, the covariation of the mean flow and wave activity is quasiperiodic for this season and latitude. We will address the timescale of covariation in Figure 4 below.

Figure 2 delineates hemispheric atmospheric flows during a period in which wave activity is high and the mean flow is weak (top row) and another period in which the opposite is true (bottom row). These instances correspond to the two blue vertical lines in Figure 1. In the former (15 December 2011) the 250 hPa geopotential reveals a highly meandering westerly flow in the midlatitude and the core of the jet is split into polar and subtropical latitudes, with westerly winds significantly weakened around 45–55$^\circ$S.
Figure 3. Covariation of wave activity and the zonal mean zonal wind at 46.5°S for the months of December-January-February. The horizontal axes are the anomalies of $\langle \hat{u} \rangle$. The vertical axes are the anomalies of (left column) $\langle A \rangle$, (middle column) $B$, and (right column) $\langle A \rangle + B$. (top row) Anomalies are defined as departures from the mean seasonal cycle. (bottom row) Anomalies are defined as departures from the 3 day running mean. Note the different scale and range of axes between the top and bottom rows. The solid red lines in the right column indicate the slope of $-1$. The dashed lines show the orientation of the major axis of the fitted ellipsoid, computed from the covariance matrix. The slopes of the major axes and the correlation coefficients are indicated in each panel. Based on ERA-Interim 6-hourly data for 1979–2013.

In the latter (16 January 2012) eddies are relatively weak and the jet exhibits a single, enhanced core around 50°S, not lowercase s. The two states exemplify the opposing phases of midlatitude eddy-mean flow vacillation captured by the covariation of barotropic wave activity and zonal mean zonal wind in Figure 1 (third panel).

Figure 3 summarizes the covariation of the barotropic zonal mean flow and FAWA at 46.5°S for the month of December through February over 35 years (1979–2013) in scatter diagrams. In all panels the horizontal axis is $\Delta \langle \hat{u} \rangle$. The vertical axes are $\Delta \langle A \rangle$ (left), $\Delta B$ (middle), and their sum (right). Each dot represents 6-hourly data. In the top row anomalies are defined as departures from the mean seasonal cycle. $\Delta \langle A \rangle$ displays a robust anticorrelation with $\Delta \langle \hat{u} \rangle$. While the data points cluster around a line, its slope is clearly steeper than $-1$, meaning that $\langle A \rangle$ varies more than $\langle \hat{u} \rangle$. The surface wave activity anomaly, $\Delta B$, on the other hand, shows a positive correlation with $\Delta \langle \hat{u} \rangle$ because the sign of $B$ is opposite of $\langle A \rangle$'s. However, the very gentle slope suggests that $B$ only slightly offsets $\langle A \rangle$, consistent with Figure 1 (third panel). As a result, ($\Delta \langle \hat{u} \rangle$, $\Delta \langle A \rangle + B$) clusters around an axis whose slope is slightly closer to $-1$ (indicated by the solid red line in the right column) but still significantly steeper. Since (10) would place the data points exactly on the red line, the scatter and deviation from the $-1$ slope suggest that effects of nonconservative processes are significant. Nevertheless, the strong anticorrelation between the barotropic flow and wave activity reveals a largely adiabatic nature of the midlatitude eddy-mean flow interaction during the austral summer.

In the bottom row, anomalies are defined as departures from the 3 day running mean. These high-frequency transients in FAWA and the mean flow exhibit much smaller variance (note the different scales of the axes) but a slope closer to $-1$ (bottom right). Notice that the change in the slope is more notable in the left column than in the middle column, suggesting that the nonconservative term $\langle \dot{A} \rangle$ in (9) is relatively small at high frequencies but not $\langle \dot{B} \rangle$.

Figure 4 (first and second panels) shows power spectra of $\langle \hat{u} \rangle$ and $\langle A \rangle + B$ as functions of frequency and latitude for 4 months of austral summer (December–March). The spectrum of $\langle \hat{u} \rangle$ is strongest at low frequency (less than 0.02 cpd) and at the flanks of the jet but minimal near the jet axis ($\sim$48°S).
Figure 4. Spectral analysis for four austral summer months (December–March) as functions of frequency (0.0167–0.25 cpd) and latitude (35°–65°S). (first panel) Power spectrum for $\langle \overset{\cdot}{u} \rangle$. Contour interval is 4 m$^2$. (second panel) Same as Figure 4 (first panel) but for $\langle A \rangle + B$. (third panel) Cospectra of $\langle \overset{\cdot}{u} \rangle$ and $\langle A \rangle + B$. Contour interval is 2 m$^2$. (fourth panel) Same as Figure 4 (third panel) but for coherence squared. Contour interval is 0.05. Zero contours are highlighted in white. Based on the ERA-Interim reanalysis (1979–2013).

This is consistent with our current understanding of the Southern Annular Mode (SAM) [Hartmann and Lo, 1998; Lorenz and Hartmann, 2001; Gerber et al., 2008]: the leading variability of the zonal mean zonal wind is a slow meridional fluctuation of the jet axis. The persistency of the SAM at low frequency is believed to result from a “positive eddy feedback,” [Feldstein and Lee, 1998; Robinson, 2000; Lorenz and Hartmann, 2001; Nie et al., 2014]. The power spectrum of FAWA is quite distinct, however. Although low-frequency peaks do appear at the flanks of the jet, FAWA is marked by a very robust periodicity around 0.03–0.05 cpd (20–30 days) between 40 and 50°S. This corresponds to the BAM found in the leading pattern of eddy kinetic energy [Thompson and Woodworth, 2014; Thompson and Barnes, 2014], and here a peak power appears at the period of 25 days. A corresponding peak is not visible in the spectrum of the zonal mean wind (Figure 4, first panel), but their cospectrum (the amplitude of the real part of the complex cross spectrum), a measure of the shared variance between FAWA and the mean flow, shows a peak at 0.04 cpd near the jet axis (Figure 4, third panel). Coherence-squared spectrum (Figure 4, fourth panel), very similar to the square of correlation coefficient, clearly identifies an enhanced correlation between FAWA and the zonal mean wind between 42–52°S and 0.04–0.07 cpd (BAM) in addition to the low-frequency peaks at the flanks of the jet (SAM). It also increases significantly at high frequency, consistent with Figure 3. (The improvement of correlation at high frequency is not captured by the cospectrum because the variances of $\langle \overset{\cdot}{u} \rangle$ and $\langle A \rangle + B$ both decrease as frequency increases.)

To understand the distinct spectra for $\langle \overset{\cdot}{u} \rangle$ and $\langle A \rangle + B$ in the midlatitudes, we show in Figure 5 power spectra of the tendency and flux terms in (4), (7), and (8) at 46.5°S for the same period as Figure 4. Figure 5 (top left) shows the power spectra of the first two terms on the right-hand side of (7) and of their sum. The convergence of the vertically integrated momentum flux (blue) has a generally broad spectrum that peaks around 0.15 cpd (~7 days) although the spectrum also contains some small peaks. The meridional eddy heat flux contribution (red) has a distinctive peak around 0.04 cpd (~25 days), at which the power is nearly twice as large as that of the momentum flux convergence. Crossover of the two spectra occurs around
Figure 5. (top left) Power spectra of total eddy forcing (vertically integrated PV flux, magenta) and contributions from eddy momentum flux convergence (blue) and from low-level meridional eddy heat flux (red). (top right) Power spectra of eddy momentum flux convergence (blue), the zonal mean zonal flow tendency \(\langle \dot{\mathcal{A}} \rangle/\langle \dot{\mathcal{U}} \rangle\) (black), and total wave activity tendency \(\langle \dot{\mathcal{A}} \rangle + \langle \dot{\mathcal{B}} \rangle\) (dashed black). (bottom left) Power spectra of eddy forcing from low-level meridional heat flux (red) and surface wave activity tendency \(\langle \dot{\mathcal{B}} \rangle\) (black). (bottom right) Power spectra of total eddy forcing (magenta) and total wave activity tendency (black). Analysis is performed at 46.5°S for the same period as Figure 4. See text for details.

0.06 cpd (\(\sim\)17 days). Thus, the predominant contribution to the total eddy forcing (the vertically integrated PV flux, magenta) comes from the momentum flux convergence in the high-frequency range and from the heat flux in the low-frequency range. Figure 5 (top right) compares the spectra of the same momentum flux convergence (blue) and the tendency of barotropic flow (black). The two spectra share similar shapes, but the latter has significantly lower power at frequency lower than 0.25 cpd (\(\sim\)4 days), suggesting the increasing role of friction represented by \(\langle \dot{\mathcal{B}} \rangle\) in (8). Figure 5 (bottom left) compares the spectra of the low-level meridional eddy heat flux term and of the tendency of surface wave activity. At low frequency the former utterly dominates the latter, suggesting that the predominant balance in (4) is between the two right-hand side terms:

\[
\frac{\partial}{\partial t} \left( \langle \mathcal{A} \rangle + \langle \dot{\mathcal{B}} \rangle \right) \approx \frac{f v' \theta'}{H(\partial \theta/\partial z)} \bigg|_{z=0} + \langle \mathcal{A} \rangle .
\]  

(11)

Notice that since \(\mathcal{B} \leq 0\), a positive \(\dot{\mathcal{B}}\) corresponds to a damping of surface wave activity. Also, since both \(\mathcal{B}\) and the meridional eddy heat flux term in (4) have the surface static stability in the denominator, the ratio of the tendency and flux terms is unaffected by the uncertainty in the surface static stability. If the left-hand side of (11) is modeled as a linear damping of \(\mathcal{B}\), the average damping time is found to be about 0.9 day, consistent with the damping time of surface temperature estimated by Swanson and Pierrehumbert [1997] over the Pacific storm track (1 day). Blanco-Fuentes and Zurita-Gotor [2011] also demonstrate a strong influence of surface thermal damping in the low-frequency variability of baroclinicity in the Southern Hemisphere midlatitude. Then by adding (4) and (7) and using (11), one obtains

\[
\frac{\partial}{\partial t} \left( \langle \mathcal{A} \rangle + \langle \dot{\mathcal{B}} \rangle \right) \approx \frac{f v' \theta'}{H(\partial \theta/\partial z)} \bigg|_{z=0} + \langle \mathcal{A} \rangle + \langle \mathcal{B} \rangle .
\]  

(12)

Figure 5 (bottom right) compares the spectra of this tendency term and the sum of the first two terms on the right-hand side. The two spectra agree well except for 0.05–0.3 cpd (3–20 days) and below 0.03 cpd (> 30 days), where the tendency has smaller power than the eddy forcing due to \(\langle \dot{\mathcal{A}} \rangle\). The source of the discrepancy below 0.03 cpd is believed to be radiative damping whose timescale in the troposphere is several weeks, whereas at 0.05–0.3 cpd it is more likely due to mixing driven by large-scale circulation.
Compared to the tendency of $\langle \bar{u} \rangle$ in Figure 5 (top right), the wave activity tendency has more power at low frequencies due to the additional heat flux contribution (compare (12) with (8); also compare the solid and dashed black curves in Figure 5, top right). This explains the generally lower power and the lack of peak around 0.04 cpd in the spectrum of the zonal mean zonal wind at this latitude as seen in Figure 4. This is also likely the main cause of the scatter and departure from the $-1$ slope in Figure 3 (third panel): since the greater tendency of FAWA in the frequency range of BAM translates to a greater variance of FAWA than the zonal mean flow, the $\Delta \langle u \rangle - \Delta \langle A \rangle$ slope becomes steeper than $-1$.

4. Summary

We have formulated the midlatitude eddy-mean flow interaction based on the latitude-by-latitude budget of barotropic FAWA, summarized in three coupled equations for the interior wave activity, surface wave activity, and zonal mean zonal flow. Since the wave activity budget can be analyzed locally at each latitude, it complements the energy budget Lorenz [1955] which requires the domain integral to close it. The analysis is also instantaneous and therefore deviates from the more traditional empirical orthogonal function approach [Thompson and Wallace, 2000; Wallace, 2000], which aims to extract patterns of flow variation from a hemispheric time series of data.

Applications to the midlatitude austral summer revealed that

1. First-order dynamics of the tropospheric eddy-mean flow interaction is quasi-adiabatic with antiphase covariation of FAWA and the zonal mean zonal flow.
2. A very robust 25 day periodicity exists in FAWA in the latitude band from 40 to 50°S, during the Southern Hemisphere summer, but the zonal mean zonal flow, while covarying with FAWA, holds a distinct spectral shape and does not exhibit a peak at 25 days. This is consistent with the BAM identified by Thompson and Woodworth [2014] and Thompson and Barnes [2014].
3. The primary reasons for this discrepancy are (a) the distinct spectra between the interior momentum flux convergence and the low-level meridional eddy heat flux and (b) a strong thermal damping of surface wave activity.

While the barotropic zonal mean wind responds primarily to the interior eddy momentum flux convergence, FAWA also depends on the meridional eddy heat flux due to strong thermal damping at the surface. It is the meridional eddy heat flux, whose power spectrum holds a distinctive peak around 25 days, that is the primary driver of the oscillation of FAWA in the frequency range of BAM. There are some uncertainties in the surface static stability that affects surface FAWA $B$ and meridional eddy heat flux term in (4). Since in the maritime boundary layer static stability tends to be weaker toward the surface, it is possible that the $1$ km vertical resolution used in this study overestimates low-level static stability. However, in that case, the true value of the meridional eddy heat flux term in (12) will be greater, and none of the above points raised above will be affected qualitatively.

We are yet to explain the spectral shape of the low-level meridional heat flux or why it differs so much from the spectrum of the interior eddy momentum flux convergence. Is it that both fluxes are driven by the same large-scale dynamics but their modulation mechanisms are different, or is it that the strong surface thermal damping also alters the timescale of heat flux modulation, or is there something fundamentally different about the dynamics that governs the two fluxes? These questions will be addressed in a subsequent paper.

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