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ABSTRACT

The relationship between the generation of African easterly waves and instability growing in regions with reversed potential vorticity gradients is studied using a regional climate model. Results indicate that the convective generation of potential vorticity (PV) due to the meridional and vertical gradients of diabatic heating in the upper and lower troposphere causes a vertically elongated PV anomaly on the southern flank of the African easterly jet. This PV maximum at 9°N in the midtroposphere, together with a PV minimum near 15°N at lower levels because of dry convection over the Sahara, reverses the meridional PV gradient between 9° and 15°N, which suggests that the zonal flow may be unstable in this region. Analysis of the seasonal mean Eliassen–Palm flux for African waves indicates that wave energy generated convectively through baroclinic overturning in the upper troposphere propagates downward and triggers barotropic conversions south of the jet and baroclinic conversions below and north of the jet.

The barotropic conversion of the jet initiates primarily outside of the region of strengthened reversed potential vorticity (q) gradients, suggesting that this barotropic conversion is a result of convectively induced eddies extracting energy from the zonal flow rather than the release of zonal kinetic energy to the waves in the unstable region. In contrast, the residual barotropic conversion occurs inside the region of reversed q gradients during the waves’ decaying stage when ITCZ convection weakens. The baroclinic instability in the unstable region becomes distinguishable from that due to surface temperature gradients when the surface heat flux is weak, a condition under which the African easterly jet better acts as an internal jet. Thus, this analysis indicates that the shear instability of the jet occurs to sustain the waves at the decaying stage rather than to initiate the waves, since it does not appear strong enough to reenergize the waves.

1. Introduction

African easterly waves originate over eastern and central Africa and propagate westward across the Atlantic. Their periods range from 3 to 5 days, over West Africa and the tropical Atlantic near 10°–15°N, to 6–9 days for waves mainly located north of 15°N (Diedhiou et al. 1998). They are known progenitors of tropical storms and hurricanes (Erikson 1963; Frank 1970).

African easterly waves have been studied using various numerical approaches and observational analyses.

Many investigations associate the 3–5-day waves with hydrodynamic instability of the African easterly jet, which forms near 600–700 hPa in the same region and at the same time of year as the waves (Burpee 1972; Rennick 1976; Simmons 1977; Kwon 1989; Thorncroft and Hoskins 1994a,b). These studies find that the Charney–Stern necessary condition for instability, that is, a reversed meridional gradient of potential vorticity (PV), is met in the vicinity of the jet’s strong wind shears, suggesting that the jet may be hydrodynamically unstable (Charney and Stern 1962; Eliassen 1983).

Other investigations into the cause of the easterly waves focus on the role of diabatic heating within the ITCZ (Holton 1971; Estoque and Lin 1977). Schubert et al. (1991) show that reversed meridional PV gradients can be generated in the midtroposphere by concentrated convective heating alone, in the absence of a background jet. Ferreira and Schubert (1997) study barotropic instability during ITCZ breakdown using a
nonlinear shallow-water model on a sphere. They find that an unstable mean flow can be produced by ITCZ convection in just a few days, suggesting that the ITCZ produces favorable conditions for its own instability and breakdown. Hsieh and Cook (2005, hereafter HC05) find that the generation of African easterly waves is more closely associated with convective heating within the ITCZ than with the strength of the African easterly jet.

Recently, an observational analysis by Kiladis et al. (2006) shows a slight phase lead of adiabatic forcing ahead of vertical motion and convection, suggesting that the convection within African waves is initiated by dynamical forcing. Mekonnen et al. (2006) analyze the European Centre for Medium-Range Weather Forecasts (ECMWF) reanalysis and satellite observations and find that convection triggered over central and eastern Africa has a role in initiating African waves. Moreover, Hall et al. (2006) study the linear normal modes of the African easterly jet and suggest that barotropic–baroclinic instability alone cannot explain the initiation and intermittence of African easterly waves, and a finite-amplitude initial perturbation is required. Hsieh and Cook (2007, hereafter HC07) find that the generation of active waves usually results from the nearly in-phase evolution of baroclinic and barotropic conversions, which are usually associated with significant rainfall over Africa.

In summary, most early studies suggest that the dynamical instabilities of the African easterly jet initiate African waves that then induce convection, while more recent studies suggest that convection causes the African waves and subsequently couples with the waves, which implies that the convection becomes phase locked with the waves. It is still unclear whether the adiabatic forcing of the jet instability induces convection that then reinforces the waves so that they can be amplified to the point that they are observable. Perhaps the jet instability alone cannot induce significant convection. Instead, the convection itself induces the waves and the waves then concentrate the convection which, in turn, enhances the waves.

The aim of this study is to further investigate how the instability of the jet and convection are involved in the generation process of African easterly waves, which may provide another way to understand whether or not Charney–Stern instability theory plays a role in initiating African waves. Or, perhaps, results of this study can also be used to evaluate if the Charney–Stern theorem is applicable to the context of African easterly waves or not. A regional climate model is analyzed to investigate the relationship between the instability of the African easterly jet and the generation of African waves through examining the mechanism of reversing PV gradients and the implied instability of the zonal flow.

The relevant literature and principles are reviewed in section 2 and, in section 3, the numerical simulations and model validation are described and presented. Section 4 explains the analysis methods used for examining the PV budget, wave–mean flow interaction, and instabilities related to PV gradient reversals. Section 5 shows the analysis results. Section 6 contains a summary and the conclusions.

2. Background

As mentioned above, African easterly wave generation is often related to the instability of the African easterly jet through a reversal of the meridional PV gradients in association with the Charney–Stern instability criterion (Charney and Stern 1962). Under the quasigeostrophic assumption, conservation of PV is approximately equivalent to the conservation of quasigeostrophic potential vorticity $q$ on pressure surfaces. Therefore, PV gradient reversals can be equivalently represented as $q$ gradient reversals as follows:

$$\frac{\partial (PV)}{\partial y} = \frac{\partial q}{\partial y} = \frac{\partial}{\partial y} \left[ f + \nabla^2 \psi + \frac{\partial}{\partial p} \left( \frac{p f_0^2}{R S_p} \frac{\partial \psi}{\partial p} \right) \right], \quad (1)$$

where $PV = -g \partial \theta / (\partial p)(s_o + f)$ is potential vorticity evaluated on isentropic surfaces, $s_o$ is relative vorticity on isentropic surfaces, $\theta$ is potential temperature, $f_0$ is the planetary vorticity at the specific latitude, $\psi$ is the geostrophic streamfunction on pressure surfaces, $p$ is pressure, $R$ is the gas constant, and $S_p \equiv \left[-(7/6)\partial \theta / (\partial p)\right]$ is the static stability. Overbars denote averaging over longitude and time. The right-hand side (rhs) of Eq. (1) includes the gradients of planetary vorticity, relative vorticity, and stretching vorticity due to the weak convergence in quasigeostrophic flow, respectively. According to the Charney–Stern necessary (but not sufficient) condition, if the zonal flow is unstable, then it requires

$$\frac{\partial q}{\partial y} = \beta - \frac{\partial^2 \psi}{\partial y^2} - \frac{\partial}{\partial p} \left( \frac{p f_0^2}{R S_p} \frac{\partial \psi}{\partial p} \right) < 0 \quad (2)$$

somewhere in the domain (Charney and Stern 1962). In Eq. (2), $\beta = df / (dy)$ is the meridional gradient of planetary vorticity, and $u$ is zonal velocity.

The second term on the rhs of Eq. (2) expresses the role of the horizontal wind shear in establishing the meridional gradients of $q \left[ \psi / (\partial y)^2 \right]$. The third term is obtained by using $u = -\partial \psi / (\partial y)$. Thus,
meridional gradients of quasigeostrophic potential vorticity $\mathbf{Q}_y$ can be reversed through a northward decrease (increase) of easterly (westerly) shear in the mean flow, where $\frac{\partial^2 \mathbf{u}}{(\partial y)^2} > \beta > 0$, or through an upward decrease (downward increase) of pressure-weighted vertical shears in the mean flow where $\frac{\partial \mathbf{u}}{\partial \phi} \left[ \frac{p^2 f^2}{P_1} \right] > \beta > 0$, or by a combination of the two.

If the necessary condition for instability is met through the horizontal shear, perturbations can grow through barotropic energy conversions. If vertical shear causes the instability criterion to be satisfied, then baroclinic conversions may support wave growth at the expense of the zonal mean flow’s energy. In the tropics, the larger values of $\beta$ tend to stabilize the flow. However, the effect of ITCZ convective heating can reverse meridional gradients of $\mathbf{q}$ and induce the horizontal and vertical wind shear without the existence of a background jet (Schubert et al. 1991).

3. Numerical simulations and model validation

Output from a regional model is analyzed to investigate the relationship between PV gradient reversals and the instabilities of the zonal mean flow over Africa. As detailed in HC05 and HC07, the regional climate model is a modified version of fifth-generation Pennsylvania State University–National Center for Atmospheric Research Mesoscale Model (MM5) that realistically captures the climatology of northern Africa and the tropical Atlantic. The model domain extends from 6°S to 45°N and 40°E to 85°W with horizontal grid spacing of 80 km. The model uses 24 vertical $\sigma$ levels from the surface to 25 hPa. A seasonal simulation from 15 May to 14 September was carried out by updating the lateral boundary conditions on winds, temperature, and moisture every 12 h from the National Centers for Environmental Prediction–National Center for Atmospheric Research (NCEP–NCAR) reanalysis climatology. Similarly, SST and soil moisture for the lower boundary conditions are interpolated from the observations of Shea et al. (1992) and Willmott et al. (1985), respectively. Therefore, in the present study, no African waves generated in the model are introduced from the boundary.

HC05 provided detailed information about the model setup and compared modeled and observed features of the climatology in five different model integrations. The model output presented here is the “realistic climate simulation” discussed in HC05. This simulation captures an accurate representation of the observed background climatology and the African easterly waves, including the phase speeds, wavelengths, and geographical distributions of the 3–5- and 6–9-day waves. For example, the simulated African waves with 3–5-day period originate near 20°E with wavelengths ranging from 2400 km at 20°E to 3500 km at 10°W. The simulated average phase speed ranges from 6 m s$^{-1}$ at 20°E to 10 m s$^{-1}$ at 10°W. Here we present additional model validation for fields of particular interest for the PV analysis.

Figures 1a,b display wind vectors and zonal wind contours interpolated onto the 319-K (near 600 hPa) isentrope from the simulation and the NCEP–NCAR reanalysis climatology (1949–2000), respectively, averaged over July and August. For the comparison, the model output has been interpolated onto 2.5° grid of the reanalysis, and the 319-K isentrope was defined using the two-month mean. The simulated African easterly jet is located a little farther south and is somewhat stronger than in the reanalysis. A small region of greater convergent flow occurs over the Cameroon highlands (centered near 6°N and 13°E) in the model, which makes it favorable for convection that causes significant rainfall over this region. This feature does not appear in the reanalysis, perhaps because of sparse observations in this region or the relatively coarse resolution of the reanalysis, or both.

Figures 1c,d show the simulated and reanalyzed relative vorticity distribution on the 319-K isentrope, respectively. The African easterly jet separates positive relative vorticity in the south from negative relative vorticity in the north in both the model and the reanalysis. Relative vorticity magnitudes are generally larger in the model output, but the location of relative vorticity maximum is quite similar to the reanalysis. The modeled relative vorticity gradient, zonally averaged between 10°S and 20°E, is significantly stronger than that of the NCEP–NCAR reanalysis. This suggests a stronger PV gradient reversal and thus a more barotropically unstable zonal flow in the model than that in the NCEP–NCAR reanalysis near the origin of African waves, consistent with the larger horizontal wind shear shown in the model compared with the climatology of the NCEP–NCAR reanalysis.

4. Analysis methods

a. PV budget

To quantitatively evaluate the production and destruction of PV in the wave generation region and identify processes that reverse the meridional PV gradient, an analysis of the PV budget is presented. The PV equation in isentropic coordinates is
is the Lagrangian derivative of PV on the meridional cross section excluding the zonal advection term, \( u\partial(PV)/\partial x \), which is moved to the rhs of Eq. (3) and treated as a PV production term along the streamlines of the meridional cross section. Here \( \sigma \) \( = -\partial p/\partial \theta \) represents the mass density, and \( \dot{\theta} \) \( = d\theta/(dt) \) denotes vertical velocity in isentropic coordinates. The thermodynamic equation shows that \( \dot{\theta} \) is proportional to the rate of diabatic heating \( Q: \theta = \theta(C_pT)Q \). In isentropic coordinates, \( \xi = \partial\dot{\theta}/\partial y - \partial\dot{\theta}/\partial x \) and \( \eta = \partial\dot{u}/\partial \theta \) \( - \partial\dot{u}/\partial x \) are the horizontal components of the 3D relative vorticity \( (\xi, \eta, \zeta) \), \( \nabla_\theta = [i\partial/(\partial x) + j\partial/(\partial y)]_\theta \) is the isentropic gradient operator, and \( F_r \) represents friction.

The second term on the rhs of Eq. (3) represents the convective generation of PV due to vertical gradients of diabatic heating, which are strong in the lower troposphere near the center of the ITCZ. This vertical stretching (compression) spins up (down) air columns to conserve angular momentum. The third term represents PV generation through the horizontal shear of vertical velocity that corresponds to horizontal gradients of diabatic heating, which can tilt the angular momentum vector from horizontal into vertical components. Therefore, horizontal vorticity resulting from strong vertical shear in the zonal flow can be transferred into vertical vorticity through this tilting process. The last term represents PV diffusion due to friction. The friction is parameterized in the Blackadar high-resolution model used in MM5 model (Zhang and Anthes 1982), and includes the vertical mixing effect due to convection.

b. Energy conversions

The energetics processes for the generation and transformation of African easterly waves were studied in detail by HC07. In this work, only the energy conversions directly associated with extracting energy from the zonal mean flow are discussed, which are those associated with the horizontal and vertical zonal wind shears, expressed as follows:
where \( p_s \) is surface pressure, and \( C_k \) is the full barotropic conversion associated with the horizontal and vertical wind shears of the zonal and meridional winds [see Eq. (A6) in HC07]. Square brackets indicate the zonal mean, and prime symbols denote deviations from zonal mean quantities. The overbar indicates the meridional average of the quantity. Additional details about the notation used here are explained in HC07. The dominant barotropic conversions are \( C_{k1} \) and \( C_{k2} \), the terms directly associated with the horizontal and vertical wind shears of the zonal mean flow, respectively. Under the quasigeostrophic condition, \( C_{k2} \) is usually neglected as part of the weak convection condition. Therefore \( C_{k1} \) is the main source of the barotropic instability associated with the waves growing in the unstable region, as suggested by the Charney–Stern necessary condition. However, in the tropics, the quasigeostrophic assumption is weaker and \( C_{k2} \) may be as large as \( C_{k1} \).

The other primary source of instability is the baroclinic energy conversion associated with the vertical shear of the zonal flow. This conversion is denoted \( C_{A1} \) [see Eq. (A8) in HC07], and it is the conversion of zonal to eddy available potential energy associated with the transport of heat down a temperature gradient. It is proportional to the vertical shear of the zonal mean flow when the thermal wind balance is satisfied above the atmospheric boundary layer. Therefore, the baroclinically unstable region of an internal jet should be located in the unstable region where the thermal wind balance holds. Most of the eddy available potential energy generated by \( C_{A1} \) is directly converted into the eddy kinetic energy of the waves through baroclinic overturning \( C_{pk} \) \( = -R\Delta p/(p_g)\)).

c. Eliassen–Palm flux analysis

In a quasigeostrophic system, barotropic energy conversions \( C_k \) are primarily due to \( C_{k1} \) [Eq. (5)] and are, therefore, proportional to horizontal eddy momentum fluxes down the mean momentum gradient. Similarly, the baroclinic conversion \( C_{A1} \) is a result of eddy heat fluxes down the mean temperature gradient [Eq. (6)]. These two quantities act together, as measured by the divergence of the Eliassen–Palm flux (EP flux; Eliassen and Palm 1961; Edmon et al. 1980; Andrews and McIntyre 1976). To include the ageostrophic eddy effect in the present study, the vector form of the full EP flux \( \mathbf{F} = (F_\varphi, F_p) \) in spherical coordinates is

\[
F_\varphi = r_0 \cos\varphi \left( \frac{\partial[u]}{\partial p} \right)_{\Theta_p} - \left[ u'v' \right] \quad \text{and} \quad F_p = r_0 \cos\varphi \left( \frac{u'}{\Theta_p} - \frac{w'u'}{y} \right) - r_0 \cos\varphi \frac{\partial[(u)\cos\varphi]}{\partial y} \frac{[v'\theta']}{\Theta_p},
\]

where \( \varphi \) is latitude, \( r_0 \) is the radius of the earth, and \( \Theta_p = \partial(\varphi)/(\partial p) \) is a measure of the static stability. The terms underlined in Eqs. (7) and (8) represent ageostrophic effects, which are usually small under quasigeostrophic conditions. For quasigeostrophic flow on a beta plane (Green 1970; Edmon et al. 1980),

\[
\nabla \cdot \mathbf{F} = \frac{1}{r_0 \cos\varphi} \left( \frac{\partial(F_\varphi \cos\varphi)}{\partial \varphi} + \frac{\partial F_p}{\partial p} \right) = (r_0 \cos\varphi)[v'q'],
\]

where the quasigeostrophic eddy potential vorticity \( q' \) is defined as

\[
q' = (r_0 \cos\varphi)^{-1} \left[ \frac{\partial v'}{\partial \lambda} - \frac{\partial(u')\cos\varphi}{\partial \varphi} \right] + \frac{\partial}{\partial p} \left( \frac{\theta'}{\Theta_p} \right),
\]

where \( \lambda \) is longitude. Under quasigeostrophic conditions, the direction of \( \mathbf{F} \) is mainly determined by the relative magnitude of the eddy momentum and heat fluxes, which correspond to barotropic and baroclinic conversions, respectively, and it provides information about the propagation direction of wave energy (group velocity) in the meridional plane (Edmon et al. 1980).

Contours of \( \nabla \cdot \mathbf{F} \) yield information about interactions between eddies and the zonal mean flow. Under the quasigeostrophic condition, the EP flux divergence is related to the flux of the quasigeostrophic eddy potential vorticity as shown in Eq. (9). The divergence of \( \mathbf{F} \), together with the quasigeostrophic potential vorticity \( q' \) distribution, shows how the eddy forcing of the zonal mean flow influences the distribution of \( q' \), and, thereby, its gradient. Note that the divergence of \( \mathbf{F} \) is calculated using the full EP flux to include the ageostrophic eddy forcing on the mean flow in this study.

5. Results

a. PV budget

Figure 2a displays the simulated PV distribution averaged from 15 June to 14 September and between 10°
and 25° E, the region in which the waves are initiated. Organized convection is not significant in this region (east of 10° E), so its effect on reversing PV gradients is minimal, allowing an examination of the causes of PV gradient reversals without the influence of organized convection near the wave initiation region over East and central Africa. In this figure, as in all subsequent figures, the variables plotted are calculated using 3-hourly model output, and then averaged over the season. If an isentrope dips underground at any of these times, that location in latitude–isentrope space is excluded from the mean and indicated by the blank areas in the figures. According to Fig. 2a, in which shading marks regions with negative meridional PV gradients, there is a wide region with negative PV gradients between the PV maximum centered near 9° N and the PV minimum between 12° and 18° N that is associated with dry convection (Thorncroft and Blackburn 1999; Dickinson and Molinari 2000) in the lower troposphere. Note also the region of reversed PV gradients in the upper troposphere, centered at 6° N between 340 and 360 K.

Figure 2b shows \((v, \theta)\) streamlines in isentropic coordinates. The confluent streamlines near 7.5° N indicate the location of the ITCZ south of the simulated African easterly jet, which is centered near 13° N and 319 K (Fig. 1a). Figure 2c displays the vertical velocity \(\dot{\theta}\), which is consistent with the diabatic heating distribution, as shown in Fig. 11c in HC05. The vertical velocity maxima at 330 K near 13° and 5° N mark regions of high condensational heating. The large condensational heating near 5° N is located south of the ITCZ, which is centered near 7.5° N, in association with a strong moisture source from the south. Descending motion near the tropopause and in the midtroposphere over the Sahara is due to radiative cooling. Everywhere south of 15° N, \(\partial \dot{\theta}/\partial \theta\) is positive below 330 K and negative above.

A comparison of Figs. 2a–c shows that, in agreement with the result of Schubert et al. (1991), regions of reversed PV gradients associated with ITCZ convection are located on the poleward side of the ITCZ at low levels (9°–15° N) and on the equatorward side of the ITCZ at upper levels (4°–7° N). Thorncroft and Blackburn (1999) show that the PV distribution of the background flow with an African easterly jet but without an ITCZ does not possess vertically elongated closed contours in the midtroposphere. In contrast, the PV distribution associated with the ITCZ convective heating alone near 10° N in their study does show vertically elongated closed contours in the midtroposphere. Both the vertical elongation of the PV anomaly in Fig. 2a and its placement suggest that it is more closely associated with ITCZ convection than with the African easterly jet.

An analysis of the PV budget reveals the production mechanisms of PV that maintain the distribution, dis-
played in Fig. 2a. Considering the possibility of interaction between a cyclonic eddy and convection that may lead to a causality problem in our seasonal mean PV analysis, we have chosen our diagnosis region east of 10°E to minimize cause-and-effect problems in our time mean PV budget analysis. As long as there is no significant interaction between cyclonic eddy and convection in this region (Norquist et al. 1977), a PV budget analysis can still offer insights about which mechanism mainly contributes to the formation of the resultant PV structure. The four components of PV generation on the rhs of Eq. (3) are plotted in Figs. 3a–d. Figure 3a displays the zonal mean zonal advection of PV by the African easterly jet. (The large values in
the upper troposphere are associated with the tropical easterly jet.)

Figure 3b shows the convective generation of PV due to the vertical gradient of diabatic heating [second term on the rhs of Eq. (3)]. Between 6°N and 12°N, that is, in the vicinity of the ITCZ, there is PV production (positive values) below 325 K and destruction above due to the spinning up of air columns below the convection maximum, and the spinning down of vortices above the convection maximum. The result is negative PV gradients on the poleward side of the ITCZ at lower levels (north of 9°N) and on the equatorward side (4°–7°N) at upper levels. Again, this is consistent with Schubert et al. (1991). The region of PV destruction in the midtroposphere south of 6°N is a result of the anticyclonic relative vorticity strengthened by the positive

\[
\frac{\partial \vartheta}{\partial y} \frac{\partial \vartheta}{\partial x}
\]

over the Cameroon highlands (Fig. 1a). Thus, there is a large positive meridional PV gradient maximum between 315 and 320 K in the midtroposphere (Fig. 2a). PV generation between 18° and 21°N, which causes the negative PV gradient north of 21°N in the midtroposphere (Fig. 2a), occurs because of a maximum in the descending motion (the effect of the vertical gradient of diabatic cooling) near 325 K (Figs. 2b,c).

PV generation, in association with horizontal gradients of the vertical motion \(\frac{\partial \vartheta}{\partial x}\) and \(\frac{\partial \vartheta}{\partial y}\) \[i.e., the tilting effect denoted by the third term on the rhs of Eq. (3)], is displayed in Fig. 3c. The maximum of PV generation at 335 K and 7°N is caused by negative meridional gradients of ascending motion in that region (Fig. 2c). Here, horizontal vorticity pointing toward the equator that results from the vertical wind shear of the easterly zonal mean flow at upper levels just south of the African easterly jet (see Fig. 3a in HC05) is transferred into cyclonic relative vorticity near 7°N at upper levels. The PV generation through the tilting effect is stronger than the destruction of PV because of the compression of air columns at 335 K near the ITCZ, and thereby it contributes to the vertically elongated positive PV anomaly shown in Fig. 2a.

Figure 3d shows the diffusion of PV due to friction [the last term of Eq. (3)]. Negative PV diffusion between 7° and 10°N is a consequence of the PV produced between 6° and 12°N through the vertical and the horizontal gradients of convective heating \[\frac{\partial \vartheta}{\partial \vartheta}, \frac{\partial \vartheta}{\partial x},\] and \(\frac{\partial \vartheta}{\partial y}\) \], as illustrated in Figs. 3b,c. In contrast, positive PV diffusion occurring at 6°N is due to the southward diffusion of PV generated north of 6°N to the region south of 6°N with decreasing PV (anticyclonic flow), as shown in Fig. 3b. To the north, the background PV is larger because of larger planetary vorticity. Thus, more of the convectively generated PV is diffused equatorward. Note that the main PV diffusion is located in the middle and upper troposphere, indicating that it is the horizontal friction that dominates the diffusion of PV rather than the vertical friction, which is mainly confined to the boundary layer. If the PV maximum is caused by a free relative vorticity due to the horizontal shear of the jet, then PV diffusion near the midtroposphere should be the dominant factor in the PV budget for the formation of this PV anomaly in the midtroposphere.

Figures 3e,f show the left-hand side (lhs) of Eq. (3) calculated in two ways. Figure 3e shows the sum of the four components on the rhs of Eq. (3) plotted in Figs. 3a–d, and Fig. 3f is an explicit calculation of \(d(PV)/dt\) from each 3-h model output using Eq. (4). The two separate calculations were performed to check the PV budget, and they are in good agreement. Comparing Figs. 3e,f with Fig. 3b suggests that the vertical gradient of convective heating is the dominant factor of causing PV maximum near 9°N in the midtroposphere and, thereby, the PV gradient reversal in the vicinity. The diffusion of convective PV generation through positive

\[
\frac{\partial \vartheta}{\partial \vartheta} \frac{\partial \vartheta}{\partial x}
\]

located near 8°N in Fig. 3b slightly shifts the net PV production maximum southward to 7.5°N (Fig. 3e).

In summary, the PV budget analysis associates the PV maximum near 9°N and the resulting PV gradient reversal in the midtroposphere with convective heating south of the African easterly jet. Without the concentrated ITCZ convective heating, PV gradients in the midtroposphere can only be weakly reversed near the jet core, resulting from the strong surface heating over the Sahara (Thorncroft and Blackburn 1999). According to the above analysis, the mean PV structure results from two climatological features, ITCZ and dry convection due to strong Saharan surface heating, which lead to reversed PV gradients north of the positive PV anomaly in the midtroposphere and reversed PV gradients south of the negative PV anomaly at lower levels north of the jet (see Fig. 2a), respectively.

In addition to the above PV budget analysis, another way of studying the instability of a zonal current is to examine reversals of quasigeostrophic potential vorticity gradients \(q_y\) on pressure surfaces, as discussed in Eq. (2). To demonstrate how the reversal of \(q_y\) is associated with horizontal and vertical shears of the mean flow, the zonal mean flow between 10° and 25°E, along with the reversing effects of horizontal and vertical curvatures of the mean flow, are displayed in Fig. 4. As shown in Fig. 4b, the southward tilting of the zero contours with increasing height between 800 and 400 hPa from 12° to 9°N is in agreement with the vanishing of the PV gradients shown in Fig. 2a. This suggests that \(q_y\) is mainly reversed by the horizontal curvature of the zonal flow

\[
\frac{\partial^2 \varpi}{\partial y^2}
\]

with a vertically elongated struc-
tecture, which primarily results from the horizontal shear closely associated with the ITCZ forcing, as also noted and shown by Schubert et al. (1991) and Thorncroft and Blackburn (1999).

On the other hand, the effect of reversing $q_y$ from the vertical curvature of the mean flow $\beta = q_y \frac{\partial}{\partial p} \left[ \frac{1}{(R S_p)} \frac{\partial}{\partial z} (p f_p) \right]$, as shown in Fig. 4c, suggests a weaker contribution to reversing $q_y$ south of the jet (14°N). In addition, the vertical curvature of the mean flow near and north of the jet, which mainly depends on the strength of the African easterly jet, reverses $q_y$ north of the jet. Therefore, the $q_y$ maximum that reverses $q_y$ south of the jet is mostly caused by horizontal shears being closely associated with the ITCZ forcing, while $q_y$ north of the jet is reversed by both the vertical and horizontal shears of the African easterly jet, consistent with the result of the PV analysis in Figs. 2 and 3.

b. Eliassen–Palm flux analysis

Since the results in Figs. 3 and 4 have shown the main cause of the PV anomaly that reverses the PV gradient east of 10°E is the convective heating within the ITCZ, there is no need to further exclude the effect of organized convection because convection is found to be an important triggering mechanism anyway. To include the eddy effects on the zonal mean flow over West Africa, the zonal average now is taken between 10°W and 25°E, which may offer information about the main wave–mean flow interaction of the African easterly jet with its maximum strength over West Africa.

To focus more on the interactions between African waves and the zonal flow, the seasonal mean EP flux and its divergence for the 3–5-day wave disturbances are shown in Fig. 5. EP flux divergence between 10° and 15°N indicates that the 3–5-day waves decelerate the easterly flow in the unstable region ($\eta_y < 0$ in Fig. 4b). Two regions of EP flux convergence are centered on 8° and 18°N with $\eta_y > 0$, suggesting a net acceleration of the easterly flow on the flanks of the jet through the 3–5-day eddy forcing. A sign of $\eta_y$ opposite to $\nabla \cdot F$ indicates a downgradient flux of eddy potential vorticity [Eq. (9)], which decreases the zonal mean gradient of potential vorticity $\eta_y$ and thus the meridional shear of the jet. Thus, the 3–5-day waves stabilize the easterly zonal mean flow through a relaxation of $\eta_y$ in the seasonal mean.

The downward-pointing vectors at upper levels between 9° and 12°N indicate that the wave energy generated through convection associated with the 3–5-day waves propagates downward on the southern flank of the jet. This downward flux, in association with the meridional eddy heat flux and the vertical eddy momentum flux [Eq. (8)], triggers barotropic conversions in
the midtroposphere south of 12°N, as shown by the vectors gradually turning from a vertical to a horizontal orientation. The northward- and downward-pointing vectors north of 12°N (the jet is centered at 600 hPa near 13°N) show that the wave energy generated through barotropic and baroclinic conversions propagates toward the surface. On both sides of the jet, the downward-pointing vectors below 700 hPa point away from the EP flux divergence center near 12°N, suggesting that the downward-propagating waves extract energy from the easterly zonal flow in this region.

c. The instability of the African easterly jet

The relationship between the generation of African waves and the instability of the African easterly jet is examined here through the analysis of energy conversions of the mean flow. In contrast to studying the shear instability of simplified zonal flows (e.g., Lipps 1970; Dickinson and Clare 1973), some stability properties for a more realistic flow can be obtained by analyzing the energetics of the system since any instability must have certain energy sources for it to grow. That is, amplifying solutions for perturbed quasigeostrophic flows usually occur in unstable regions (Dickinson and Clare 1973; Dickinson 1973), suggesting that positive barotropic or baroclinic conversions, or both, usually occur in the region with reversed PV gradients. Therefore, only energy conversions related to the zonal mean flow are discussed here (see section 4d). The full energetics for the wave generation was analyzed by HC07.

Before the discussion of the related energy conversions, the major wave activity events, along with the precipitation, are introduced here. Figure 6 displays the simulated daily mean precipitation and perturbation geopotential heights with perturbation wind vectors for African waves (3–5-day periods) at 725 hPa on 24 June, 15 July, 18–21 July, 3 September, and 6 September. On 24 July (Fig. 6a), the waves possess a pronounced southeast–northwest tilt north of the jet and a slight southwest–northeast tilt south of the jet (around 13°N), indicating stronger positive barotropic conversions north of the jet than south of the jet. Some significant precipitation falls between 5° and 9°N near the southern edge of wave trough (negative geopotential height perturbations).

Figure 6b shows significant negative geopotential height perturbations associated with concentrated precipitation near 8°E–10°N in the wave trough. The perturbed winds of this cyclonic eddy extract energy from the zonal mean flow (barotropic conversions) between 9° and 12°N near 10°–15°E (not shown), suggesting the initiation of a significant wave disturbance through concentrated convection. This convectively induced disturbance continues to grow and interact with the convection. (The evolution of moist convection on 16 and 17 July is shown in Figs. 6b,c in HC07.)

Figure 6c displays the enhanced wave disturbances and convection on 18 July. As noted by HC07, both the deep moist convection and the cyclonic vorticity induced through barotropic conversions contribute to geopotential height perturbations in the midtroposphere, which in turn may further enhance deep moist convection in the wave trough. The advection of significant rainfall by the strong cyclonic flow near 14°E demonstrates the gradual organization of precipitation around the wave trough. When the waves are strongly coupled to moist convection, heavy concentrated rainfall develops and migrates with the trough, as shown in Fig. 6d. The resulting intense precipitation (over 70 mm day⁻¹ near 0°E on 20 July) rapidly decays when the wave trough reaches 5°W on 21 July, as shown in Figs. 6e,f.

Figure 6g shows the cyclonic perturbation winds and concentrated precipitation associated with the wave trough (12°N, 9°W) on 3 September. The somewhat wavy band of precipitation extends westward to the coast, a prominent signature of African easterly waves. As shown in Fig. 6h, the undulating precipitation strip extends from the west coast of Africa to the wave trough near 5°W on 6 September.

In summary, Fig. 6 displays the horizontal structures of the most significant African waves, along with the corresponding rainfall in the model, to investigate the association between African waves and precipitation. The figure demonstrates that African waves may possess different properties with different generation mechanisms. For instance, the waves with significant northwest–southeast tilting axes are caused by barotropic conversions north of the jet due to the strong dry convection near the Sahara (Fig. 6a). The gradual formation of phase locking between the waves and convection is shown in Figs. 6b–f. African easterly waves can also be generated by convectively induced barotropic conversions by ITCZ without baroclinic instability at lower levels north of the jet (Figs. 6g–h).

Figure 7a displays the evolution of the area mean baroclinic overturning term Cpk and the barotropic conversion term Ck vertically integrated from the surface to 125 hPa. The domain for calculating the area means of these energy conversions extends from 10° to 25°W in longitude and 5° to 25°N in latitude, slightly larger than the analysis domain in HC07. As shown in Fig. 7a, full barotropic conversions tend to gradually increase with baroclinic overturning in a nearly in-phase oscillatory mode. The phase locking between barotropic conversions and baroclinic overturning starting around 18–21 July gradually leads to organized precipitation mi-
FIG. 6. Daily mean precipitation and 3–5-day wave disturbances shown with perturbation wind vectors and filtered geopotential heights at 725 hPa with a contour interval of 1 m for (a) 24 Jun, (b) 15 Jul, (c) 18 Jul, (d) 19 Jul, (e) 20 Jul, (f) 21 Jul, (g) 3 Sep, and (h) 6 Sep. Vector scale is indicated in m s$^{-1}$. Precipitation is indicated by shading with an interval of 10 mm day$^{-1}$. 
grating with the waves over West Africa, as shown in Figs. 6c–e, indicating strong wave–convection coupling through the resonance between the baroclinic overturning and barotropic conversions (HC07). In contrast, the nearly in-phase oscillations of barotropic conversions and baroclinic overturning in late June and early September do not cause such amplification. Near the end of August, $C_k$ gradually increases and oscillates with $C_{pk}$ with a slight phase lag, suggesting that $C_k$ may be gradually induced by $C_{pk}$ because of moist convection. The increased $C_k$ may help in the formation of a more concentrated convection that can strengthen the reversed potential vorticity gradients south of the jet (e.g., 3–4 September), which in turn triggers the occurrence of a more unstable zonal flow, for example, the wave events starting around 13–17 July and 1–4 September.

A comparison between the area means of the integral barotropic conversions $C_k$ and $C_{k1}$ is shown in Fig. 7b. From late June to early July, $C_{k1}$ is not the dominant source of $C_k$, suggesting that the barotropic conversions associated with the horizontal shear are not the key mechanism for generating the unstable zonal flow on which the waves grow during this period. Near 1 July, $C_{k1}$ is nearly zero while $C_{pk}$ is still strong because of the active ITCZ convection (see Fig. 5 in HC07). This indicates that although active waves often coincide with increases in barotropic conversions, they can exist and be maintained without significant barotropic conversions of the zonal mean flow.

Fig. 7. Time series of area means of the conversion rate for (a) baroclinic overturning $C_{pk}$ (solid line) and full barotropic conversion $C_k$ (dashed line), (b) the comparison between $C_k$ (dashed line) and $C_{k1}$ (solid line), and (c) $C_{pk}$ (solid line) and $C_{A1}$ (dashed line). Unit is W m$^{-2}$. 

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In early July, baroclinic overturning $C_{pk}$ starts to decay, corresponding to a weakening of ITCZ convection. However, $C_k$ begins to increase even though $C_{k1}$ is still rather weak, which is primarily a result of significant barotropic conversions in association with the vertical wind shear ($C_{k2}$), mainly caused by strong shallow convection beneath the jet during a relatively dry period near 3 July (see Figs. 4, 5, and 7b in HC07). This suggests that active African waves can be generated without involving significant rainfall (deep moist convection).

Figure 7c shows the area means of $C_{A1}$ and $C_{pk}$ evolving with time. From mid-June to 1 July, the overall eddy available potential energy produced by $C_{A1}$ is negligible in comparison with that consumed by $C_{pk}$. This indicates that baroclinic conversion $C_{A1}$ does not contribute significantly to the growth of the waves during this time. The eddy available potential energy produced by $C_{A1}$ gradually increases and becomes the main source of $C_{pk}$ when $C_{pk}$ is weak, around 5 July. The area mean of $C_{A1}$ reaches its maximum around 20 July when the waves are strongly coupled with convection. In contrast, $C_{A1}$ appears relatively small in August and September. This suggests that the baroclinic conversion $C_{A1}$ does not play an important role for the generation of these modeled waves in late summer, although the small magnitude of $C_{A1}$ leads to relatively weak wave activity during this period.

d. Reversal of potential vorticity gradients

The above analysis shows that energy conversions from sources other than $C_{k1}$ and $C_{A1}$ also make substantial contributions to the generation and maintenance of African waves, for example, baroclinic overturning through convective heating. However, it is still not clear how the unstable zonal mean flow is associated with the wave generation. In this section, the reversal of quasigeostrophic potential vorticity gradients ($q_y$) is studied along with barotropic and baroclinic conversions ($C_{k1}$ and $C_{A1}$).

Considering that the average doubling time for eddy kinetic energy of African waves in this study is about 1 day (HC07), day-by-day energy conversions can reveal how the waves grow and change the ambient PV gradient. Therefore, we analyze the daily mean energy conversions of the mean flow $C_{k1}$ and $C_{A1}$ and quasigeostrophic potential vorticity $q$ distribution in the meridional cross section to investigate how the simulated zonal flow is destabilized.

Figures 8a,b show the daily mean distributions of $q$ and $C_{k1}$ in the meridional cross section on 23 June during a major modeled wave event. The primary barotropic conversion $C_{k1}$ is larger on the northern flank of the jet than the southern flank of the jet. As noted by Burpee (1972), the easterly eddy momentum transport is mainly poleward in June and equatorward in July, August, and September. Two barotropic conversions are mainly located outside the negative $q_y$ region on the two sides of the jet, suggesting that the barotropic conversion here is mainly not a result of Charney–Stern instability.

On 24 June, shown in Figs. 8c,d, a robust reversed $q_y$ reversal on the northern flank of the jet forms and strengthens because of the formation of a $q$ minimum between 600 and 700 hPa and a local maximum of $q$ near the surface around 18°N, which are associated with the strong dry convection. The main barotropic conversions near 18°N result from the convectively strengthened anticyclonic eddy extracting energy from a more unstable flow with strengthened $q_y < 0$ north of the jet, which is consistent with the southeast–northwest tilt of the wave axis north of the jet (Fig. 6a). In contrast, negative barotropic conversions occurring adjacent to the region with negative $q_y$, that is, the region where $q_y > 0$, suggest that eddy kinetic energy is absorbed to strengthen the zonal flow, in agreement with the stable region indicated by positive $q_y$.

Dry convection starts to decay on 25 June, corresponding to less concentrated barotropic conversions north of the jet and a relatively weak negative $q_y$. The negative barotropic conversions in the region with weak $q_y > 0$ also become weak. The barotropic conversion in the region with positive $q_y$ (stable region) around 7°N suggests that the destabilizing effect on the zonal flow caused by the convectively induced eddy is stronger than the stabilizing effect due to $q_y > 0$.

As shown in Figs. 7a,b, the barotropic conversions near 12–14 July are rather weak and gradually increase as they oscillate with baroclinic overturning, suggesting that the weak barotropic conversions ($C_{k1}$) are caused by weak eddies energized by baroclinic overturning ($C_{pk}$). These eddies are gradually intensified by $C_{pk}$ and $C_{k1}$ and generate a larger pressure perturbation that helps the formation of a more concentrated convection in the wave trough, at 8°E–10°N on 15 July (Fig. 6b). The concentrated convection may in turn strengthen the reversed meridional gradients of $q$, leading to a more unstable zonal flow.

The daily mean $q$ and the barotropic conversions on a zonal mean meridional cross section for the initiation of wave–convection coupling from 16 to 18 July are displayed in Fig. 9. As shown in Figs. 9a,b, the main barotropic conversion $C_{k1}$ is located outside the region of $q_y < 0$. This suggests that the barotropic conversions are not a result of releasing energy from the unstable zonal flow as suggested by the Charney–Stern neces-
sary condition. Instead, they are a consequence of convectively induced eddies extracting zonal kinetic energy outside the unstable region and consuming eddy kinetic energy inside the region of \(q_y < 0\). Note that the generated eddies are also maintained by the barotropic conversions from other sources, for example, \(C_{k2}\), because of shallow convection beneath the jet (see Fig. 7b).

Starting from 17 July and shown in Figs. 9c,d, ITCZ convection is concentrated along 9°N, as shown in Fig. 6c of HC07, leading to the strengthening of the \(q\) maximum centered at 10.5°N and, thereby, its meridional gradients \(q_y\). The corresponding barotropic conversion maximum increases and mainly still operates outside the unstable region. On 18 July, the barotropic conversion \(C_{k1}\) rapidly reaches a maximum while \(q\) and its gradients only slightly weaken (Figs. 7b and 9e,f). This strong barotropic conversion at 10.5°N generates robust waves with cyclonic–anticyclonic eddies accompanied by large pressure perturbations near the African easterly jet, which induces even stronger deep moist convection and organizes the precipitation through the advection of the cyclonic flow (Fig. 6c). The strong coupling between the waves and convection (including both deep moist convection and shallow convection) from 18 to 20 July maintains both the strengthened \(q_y\) and barotropic instability of the zonal mean flow (Figs. 7a,b).

The wave–convection coupling starts to break apart on 21 July (see Fig. 6f). As displayed in Figs. 10a,b, the

![Daily means averaged between 10°W and 25°E on 23 Jun](http://journals.ametsoc.org/jas/article-pdf/65/7/2130/3505090/2007jas2552_1.pdf)
The convectively induced unstable zonal flow starts to release its kinetic energy to wave disturbances when the deep convection weakens, and this relaxes $q_y$. As the coupling weakens, the $q$ maximum shifts southward, leading to a more compact unstable region between 10.5° and 14°N where the maximum barotropic conversion occurs on 22 July (Figs. 10c,d). On 23 July, as shown in Figs. 10e,f, the $q$ maximum shifts further southward to 9°N and the barotropic conversion of the zonal flow overlaps with the unstable region with $q_y < 0$. This suggests that the decaying waves are maintained in the unstable region at the expense of the zonal mean kinetic energy, consistent with the shear instability property as suggested by the previous studies (e.g., Charney and Stern 1962; Dickinson and Clare 1973). Note that from the above analyses for these case studies the wave growth and maintenance is taking place at the expense of the mean flow in unstable regions even though the waves are mainly caused by convectively induced instability at the initial stage.

Figure 11 shows the daily mean zonal mean $q$ and $C_{k1}$ cross sections on 3, 6, and 9 September. The zonal mean $q$ and its gradients on 3 September are strengthened because of deep convection south of the jet (Fig. 6g). The cyclonic wind perturbations cause positive barotropic conversions between 9° and 12°N where $q_y > 0$ and $C_{k1} > 0$.
weak negative barotropic conversions in the unstable region with $q_y < 0$ between 12° and 15°N. Disturbances are not growing in the unstable region. Rather, they are decaying to strengthen the zonal mean flow where $q_y < 0$ near the jet core. This negative barotropic conversions result from the cyclonic eddy, mainly induced by the concentrated moist convection near 12°N, interacting with the cyclonic wind shear of the jet with its core centered around 15°N on 3 September. The eddy momentum flux of the perturbed southeasterly winds south of 15°N ($-u'\upsilon' > 0$), as displayed in Fig. 6g, correlates with the weak cyclonic zonal wind shear near the jet core $[\partial u/\partial y] < 0$, causing negative barotropic conversions between 12° and 15°N.

Because of the persistent concentrated convective heating in early September, the daily mean $q$ gradients rapidly increase on 4 September (not shown), corresponding to large $C_{k1}$ values of around 0.03 W m$^{-2}$, as shown in Fig. 7b. Such convectively induced barotropic conversions continue to provide energy to the waves on 6 and 9 September. Note that the zonal mean $q$ maximum shifts southward to 9°N as the main deep moist convection moves southward and becomes more concentrated between 6° and 9°N on 9 September, leading
to more closely packed isolines of $q$, suggesting a more unstable region of $q_y < 0$ for the waves to grow and a more stable region of $q_y > 0$ to absorb the wave energy to the south of the $q$ maximum. This indicates a stronger interaction between positive $q_y$ and negative $q_y$, which may explain why more barotropic conversions now occur in the region of $q_y < 0$ because the convectively induced barotropic conversions are weaker, and barotropic instability is inhibited more effectively in the more stable region. The above results suggest that the barotropic instability of the jet occurs to sustain the waves at the decaying stage when convection is weak rather than to initiate the waves because it does not appear strong enough to reenergize the waves and lead to organized convection.

In addition to the role of barotropic conversions of
the zonal mean flow in generating African waves, baroclinic conversions in association with the occurrence of $q_y < 0$ are discussed. As displayed in Fig. 7c, the baroclinic conversion $C_{A1}$ gradually becomes more important in early and late July, and then is negligible during August and September. The daily mean $q_y$ and $C_{A1}$ from 1 to 6 July are shown in Fig. 12. The major baroclinic conversion is located near 16°N in the lower troposphere, a result of the instability from the surface with strong potential temperature gradients south of the Sahara. Instability due to strong baroclinic stratification drives the surface heat flux across mean isotherms, converting zonal available potential energy to eddy available potential energy. The eddy available potential energy is converted to eddy kinetic energy through sloping convection originated from the surface confluent zone.

A secondary baroclinic conversion maximum also exists between 700 and 800 hPa near 12°N in the unstable region from 1 to 3 July, particularly distinguishable when baroclinic conversions near the surface around 15°N are relatively weak on 2 July. [This condition is more favorable for the instability of a quasigeostrophic internal jet, as suggested by Charney and Stern (1962).]
The vertical zonal wind shear above 800 hPa near 12°N is a consequence of the thermal wind balance that leads to the formation of the African easterly jet (Cook 1999; Thorncroft and Blackburn 1999). As argued by Burpee (1972), the Charney–Stern theory is still applicable even when the meridional gradient of the surface potential temperature is not zero as long as the amplitude of the perturbation is negligible at the surface. HC07 also showed that the wave activity near the surface around 12°N is significantly weaker than that north of the jet and can be considered negligible in comparison (see Fig. 13 of HC07). Note that there are some negative barotropic conversions located around 10°N between 600 and 700 hPa where \( q_y \) > 0 on 1 and 2 July (not shown). This suggests that the African easterly jet is barotropically stable near the midtroposphere but baroclinically unstable below the jet core during this period.

Starting from 3 July, the reversed \( q_y \) weakens and baroclinic instability in the unstable region triggers stronger interactions between negative \( q_y \) and the positive surface potential temperature gradients \( \theta_y \) on the southern flank of the Sahara (Pytharoulis and Thorncroft 1999), leading to an even larger baroclinic instability due to large positive \( \theta_y \) at the surface north of the jet. From 4 to 6 July, a relatively dry period (see Fig. 5 of HC07), the reversed \( q_y \) is significantly weakened as expected, while the baroclinic instability resulting from positive \( \theta_y \) extends southward to 12°N below 900 hPa when the meridional distribution of these two baroclinic instabilities merge. On 5 and 6 July, baroclinic instability occurs outside the region of diminishing \( q_y \) < 0, indicating that the baroclinic instability due to positive surface temperature gradient is the main source of eddy available potential energy when ITCZ convection is weak (see Fig. 7a).

Figure 13 displays the daily mean zonal mean \( C_{A1} \) and \( q_y \) from 16 to 23 July, the period of active \( C_{A1} \) as shown in Fig. 7c. On 16 July, positive baroclinic conversions along with weak negative barotropic conversions in the region of \( q_y \) < 0 shown in Fig. 9b explain the baroclinically unstable zonal flow between 700 and 800 hPa around 13°N. Baroclinic instability generated from the surface outside the region of \( q_y \) < 0 is located further north, near 18°N. As shown in Figs. 9d and 13b, the convectively strengthened reversed \( q_y \) around 12°N leads to combined barotropic and baroclinic conversions of the jet on 17 July. From 18 July to 22 July, the waves are coupled with convection, which causes strong baroclinic conversions through sloping convection north of the jet in the lower troposphere, as displayed in Figs. 13c–e. The strong baroclinic conversions north of the jet result from the large southward heat flux \( (u'v' < 0) \) due to strong thermal advection across isotherms. The weak negative baroclinic conversions near 700 hPa around 12°N (around the region of \( q_y < 0 \)) are a consequence of the northward heat flux \( (u'v' > 0) \) correlating with positive vertical wind shear \( \frac{\partial\theta}{\partial y} > 0 \), which may result from cold ascending air in the wave trough or warm descending air in the ridge, both corresponding to negative baroclinic overturning \( (w'u' < 0) \) forced by positive barotropic conversions (see Figs. 9f and 10b,d,f) near the same region. On 22–23 July, baroclinic conversions mainly occur outside the unstable region, suggesting a barotropically unstable jet profile as the convection further weakens. The meridional eddy heat flux near the surface plays a major role in maintaining these waves in the lower troposphere during the waves’ decaying stage.

The above results show that the baroclinic instability of the jet contributes to the growth of African waves in the unstable region with reversed potential vorticity gradients around 12°N. However, baroclinic instability resulting from the positive surface temperature gradients plays a major role in the generation and maintenance of these waves at lower levels north of the jet, especially when the waves are coupled to convection. In this case, the cold ascending air forced by barotropic conversions south of the jet may lead to negative baroclinic conversions in the unstable region and the meridional heat flux from the surface north of the jet becomes the main driving force for these waves in the lower troposphere.

6. Summary and conclusions

African easterly waves have long been associated with the barotropic and baroclinic instability of the African easterly jet through meridional potential vorticity gradient reversals, that is, the Charney–Stern necessary condition for instability. Here we evaluate the processes of reversing PV gradients in a regional model with a realistic simulation of African easterly waves to distinguish between the following two scenarios:

1) African easterly waves are initiated by the instability of the African easterly jet through the reversal of PV gradients, and convection is induced as a result.
2) Convection strengthens the reversal of PV gradients and convectively induced instability generates African easterly waves. The waves in turn enhance and organize the convection. The shear instability of the jet occurs to sustain the waves at the decaying stage rather than to initiate the waves.

To distinguish between these two possibilities, we investigate the generation of African waves in association with the processes that cause the reversed meridional
PV gradients, which are both observed and captured in the simulation, by a close examination of the PV budget and the related instability of the zonal flow. The results are summarized as follows:

1) The waves originate over East and central Africa near a region with reversed (negative) meridional PV gradients between 9° and 15°N, and extending from 310 to 335 K (roughly 800–400 hPa). These negative PV gradients occur because there is a strong PV maximum centered near 9°N, close to the center of the ITCZ (Fig. 2a), and a PV minimum near 15°N at lower levels. The analysis demonstrates that this PV maximum is a result of convection. Specifically, the midtropospheric convection is associated with high vertical gradients of diabatic heating.
in the lower troposphere, and these are the primary cause of the PV maximum near 320 K at 9°N (Fig. 3b). Above 320 K, high horizontal gradients of vertical motion also contribute (Fig. 3c), extending the PV anomaly upward to 335 K. While the strongest net production of PV occurs near 7.5°N, the increase of planetary vorticity \( f \) with latitude causes the PV maximum to be centered at 9°N, slightly north of the maximum production region.

2) An EP flux analysis is performed using model output filtered to select 3–5-day periods to focus more precisely on interactions between the 3–5-day African easterly waves and the zonal mean flow. This diagnosis reveals that convectively generated wave energy associated with baroclinic overturning propagates downward from the upper to the midtroposphere south of the jet, triggering barotropic conversions on the southern flank of the jet and baroclinic conversions beneath and north of the jet.

3) The instability of the jet is studied along with the reversed \( q_y \) to understand whether the implied instabilities by the Charney–Stern theory initiate African easterly waves. Results indicate that at the waves’ inception, barotropic conversions occur mostly outside the region with reversed \( q_y \), indicating that the waves are convectively induced instabilities that destabilize the jet. In contrast, during the waves’ decaying stage, when convection is weaker, the residual barotropic instability occurs within the unstable region with reversed \( q_y \), suggesting the release of zonal kinetic energy from a barotropically unstable jet caused by the Charney–Stern instability.

4) Baroclinic instability, that is, the conversion of zonal to eddy available potential energy, occurs both inside and outside the region with negative \( q_y \). Inside the region with negative \( q_y \), baroclinic energy conversions are associated with vertical zonal wind shear of the jet. They are weaker than the baroclinic energy conversions outside the region of negative \( q_y \), where the meridional flow advects the surface heat flux across isotherms (strong surface potential temperature gradients). When this advection is weak, the baroclinic instability of the jet becomes more distinguishable from the baroclinic instability associated with surface potential temperature gradients, and the African easterly jet is more accurately described as an internal jet.

In summary, this study indicates that barotropic and baroclinic instabilities within a region of reversed potential vorticity gradients are generally weaker than instabilities induced by convection or caused by strong baroclinic stratification because of the pronounced surface temperature gradients of northern Africa. We conclude that the shear instability of the jet occurs to sustain the waves at the decaying stage rather than to initiate the waves that may subsequently couple with convection. This agrees with the results of Hall et al. (2006), who note that the African easterly jet will not generate African easterly waves without finite-amplitude perturbations. Here we find that the required finite-amplitude perturbations are generated or energized by other energy sources, for example, baroclinic overturning through convection. Without concentrated condensational heating and strong surface heat fluxes north of the jet, the African waves growing at the expense of zonal mean energy in the unstable region would be weak because of the weaker stretching vorticity effect due to larger Rossby number and the stronger \( \beta \) effect near the equatorial region.

Although easterly waves over Africa and elsewhere are all convectively induced instabilities, some differences still exist between them in their properties and energetics. The uniqueness of African easterly waves lies in the existence of zonal kinetic and available potential energy from the African easterly jet, which provides an extra energy source for the waves to grow if instability is triggered to extract the energy from the zonal current. It is found that the African easterly jet plays a role in setting the scale of the waves and its structure even though the instability of the jet is not the main cause of African waves.

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