Generation of African Easterly Wave Disturbances: Relationship to the African Easterly Jet

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ABSTRACT

The relationship between African easterly waves and the background climatology in which they form is studied using a regional climate model. The surface and lateral boundary conditions in the model are manipulated to modify the background climatology, especially the African easterly jet and the ITCZ, and the behavior of the waves in these different settings is evaluated.

Three climate simulations are presented, with monthly mean lateral and surface boundary conditions. One has a strong jet and a strong ITCZ, the second has a strong jet and a weak ITCZ, and the third has a weak jet and a strong ITCZ. In these simulations, the presence of wave activity is more closely associated with the concentration of the ITCZ than the strength of the African easterly jet. In particular, the simulation with a strong jet accompanied by a weak ITCZ does not produce significant wave activity, but a weak jet with a strong ITCZ has realistic wave disturbances. Both the Charney–Stern and the Fjörtoft necessary conditions are satisfied in all three simulations, suggesting that combined barotropic and baroclinic instability contributes to the generation of waves. Near the origin of the waves, meridional gradient reversals of isentropic potential vorticity result from potential vorticity anomalies generated by convective heating within the ITCZ, implying that the unstable zonal flow may be caused by cumulus convection within the ITCZ and not by shear instability associated with the jet.

Two additional simulations with 1988 lateral boundary conditions demonstrate that 3–5-day wave disturbances can be generated in the absence of the African easterly jet, but with unrealistically small wavelengths. These results suggest that African easterly waves are initiated by cumulus convection within the ITCZ, and not by barotropic instability associated with the jet.

1. Introduction

African easterly waves are an important synoptic feature of the summer climate in West Africa and the tropical Atlantic. Observations during the Global Atmospheric Research Program (GARP) Atlantic Tropical Experiment (GATE) revealed synoptic-scale disturbances with periods of 3–5 days and wavelengths of 2000–4000 km propagating to the west across West Africa and the tropical Atlantic with phase speeds of 6–8 m s$^{-1}$ (Carlson 1969; Burpee 1972; Reed et al. 1977). The waves regularly traverse the Atlantic to reach the Central and North American coasts and are associated with the formation of tropical cyclones in the western Atlantic Ocean and Caribbean Sea (Riehl 1954; Thornicroft and Hodges 2001). Frank (1970) showed that about half of the tropical cyclone formation in the Atlantic is related to African easterly waves. Avila and Pasch (1992) investigated Atlantic tropical systems during the 1991 hurricane season and found that nearly all of the eastern Pacific tropical cyclones were associated with African waves.

Diedhiou et al. (1998) distinguish similar periods in the National Centers for Environmental Prediction–National Center for Atmospheric Research (NCEP–NCAR) reanalyses and find that 3–5-day waves have average wavelengths of 2500 km and phase speeds around 8 m s$^{-1}$. They note two preferred tracks, one to the north and one to the south of the African easterly jet, and suggest that these tracks tend to merge over the Atlantic. The 6–9-day waves originate farther north than 3–5-day waves, have longer wavelengths of about 5500 km, and move westward at a lower phase speed of approximately 6 m s$^{-1}$.

Numerous studies address the causes and dynamics of African easterly waves, but outstanding questions remain. Many investigations focus on the hydrodynamic stability of the zonal flow since African easterly wave activity occurs in the vicinity of the African easterly jet. This midtropospheric feature is centered near 15°N in summer and is a consequence of the strong positive meridional temperature and geopotential

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height gradients that develop over Sahelian Africa (Thorncroft and Blackburn 1999; Cook 1999).

The relationship between the jet and the waves is investigated here in a series of synoptic and climate-mode simulations with a regional climate model. A companion paper (Hsieh and Cook 2004b, manuscript submitted to J. Atmos. Sci.) presents a thorough energetics analysis of the model output. Central to the believability of both investigations is the quality of the African easterly wave simulation in the model. After a review of the pertinent literature in section 2, the model is described (section 3) and shown to produce an excellent reproduction of the African easterly wave climatology in section 4. In section 5, the modeling results are analyzed to investigate the relationship between the jet and the waves. Section 6 is the summary and conclusions.

2. Background

Much of the theoretical work concerned with the generation of these waves focuses on the African easterly waves as an expression of hydrodynamic instability associated with meridional gradient reversal of isentropic potential vorticity (IPV). Fjortoft (1950) applied Rayleigh theory to atmospheric flows and obtained a necessary condition for barotropic instability, which requires that the zonal velocity be positively correlated with the meridional gradient of potential vorticity. Charney and Stern (1962) generalized the instability problem to three dimensions and applied it to an internal jet by including horizontal and vertical wind shear in a stratified atmosphere. They showed that the instability of such a jet requires a sign reversal of the meridional IPV gradient; that is, a disturbance cannot grow in zonal flow that has a positive meridional IPV gradient everywhere in the domain (Eliassen 1983).

Burpee (1972) used the Charney–Stern necessary condition to explain the existence of African easterly waves, even though the African easterly jet cannot be strictly classified as an internal jet. His study suggested that the generation of African easterly waves is directly related to the African easterly jet since the Charney–Stern necessary condition is satisfied near the jet. The theory relating the waves and the jet was further developed in the modeling studies of Rennick (1976) and Simmons (1977), who considered confined zonal flows and suggested that the waves grow at the expense of the zonal kinetic energy of the jet without large influences from latent heat releases.

Thorncroft and Hoskins (1994a,b) and Thorncroft (1995) used the primitive equations with the hydrostatic condition to study the linear and nonlinear behavior of easterly waves near the easterly jet. Their studies indicate that initial wave generation occurs through barotropic conversion but that later growth is supported by baroclinic conversion in a nonlinear process. They conclude that African easterly waves grow by combined barotropic and baroclinic instabilities associated with the jet, but note that the role of latent heating is important in increasing the baroclinic conversion and determining the synoptic structure of the waves.

Other investigations emphasize the role of diabatic heating in generating wave disturbances. Holton (1971) suggested that latent heat released in convection driven by frictional moisture convergence is of primary importance for waves that form along the ITCZ. Mass (1979) studied African waves using a linear model and found that convective heating dramatically increases the perturbations in the lower and upper troposphere, in agreement with observations (Burpee 1974; Reed et al. 1977). Schubert et al. (1991) investigated the potential vorticity characteristics of the ITCZ using a simple zonally symmetric balanced inviscid model. They found that concentrated convective heating can cause large IPV gradient reversals and suggested that the existence of the African easterly jet is not necessary for the waves to occur. Ferreira and Schubert (1997) used a nonlinear shallow water model on the sphere to simulate the ITCZ breakdown and found that ITCZ convection plays an important role in the origin of the unstable mean flow in which easterly waves form.

Finally, whether the waves are associated with Charney–Stern instability (the reversal of IPV gradient) is also studied. Chang (1993) studied the unstable waves under the near-neutral conditions over the Sahara and found that waves in this region occur through baroclinic energy conversions where the Charney–Stern necessary condition is not satisfied.

3. Model and simulation description

Considering some limitations of simpler models, for example, the hydrostatic assumption and simple parameterizations of latent heat release, a nonhydrostatic model with the ability to realistically capture the climatology of northern Africa and the tropical Atlantic may offer an alternative modeling approach for studying the relationship between African easterly waves and the jet. In this study, climate-mode simulations with a regional model are performed to study the relationship between the background climatology and wave activity in the absence of interannual variability. In addition, synoptic-mode experiments are conducted to simulate the summer of 1988. According to Avila and Clark (1989), the African easterly waves of 1988 were vigorous during the entire summer, with 62 tropical waves, 19 tropical depressions, and 12 tropical storms.

The fifth-generation Pennsylvania State University (PSU)–NCAR Mesoscale Model (MM5; version 3.4) (Grell et al. 1994) is used. MM5 is a regional model based on finite differencing in both space and time. Its nonhydrostatic dynamics, with 24 vertical sigma levels, allows the model to be valid for mesoscale simulations.
Here, large domains are used to position the lateral boundaries far from the area in which the waves occur and eliminate the possibility that disturbances are propagating into the domain rather than being generated by the model physics.

Three climate-mode simulations are presented. Each is 124 days in length, and a horizontal resolution of 80 km is used across the domain, shown in Fig. 1. The model is initialized at 0000 UTC on 15 May using climatological May conditions from the NCEP–NCAR reanalysis (Kalnay et al. 1996) and run through the middle of September. The first 17 days are discarded for model spinup. Lateral boundary conditions for winds, temperature, and moisture at each vertical level are also taken from the NCEP–NCAR reanalysis, updated every 12 h based on a linear interpolation between monthly means as in Vizy and Cook (2002). Thus, the lateral boundary conditions include the influence of the seasonal cycle, but not the diurnal cycle. Sea surface temperatures are prescribed and also updated every 12 h using linear interpolation of the monthly SST climatology of Shea et al. (1992).

Considerable testing and validation were performed to choose a set of lateral boundary locations, surface and lateral boundary conditions, and physical parameterizations that produce a good simulation of West Africa and the adjacent Atlantic Ocean (Vizy and Cook 2002, 2003). The radiation parameterization used is the Rapid Radiative Transfer Model (RRTM) longwave scheme (Mlawer et al. 1997). This scheme combines the cloud-radiation shortwave scheme and the simple ice explicit moisture scheme (Dudhia 1989) to determine shortwave fluxes and cloud ice. The RRTM radiation scheme was chosen because it produces a much more realistic surface radiation budget in our simulation compared to the other radiation parameterizations in the model. The Community Climate Model Version 2 (CCM2) scheme, for example, consistently produces far too little solar radiation incident at the ground because of its strong interaction with low clouds. Other parameterizations chosen are the Kain–Fritsch (1993) cumulus scheme, the high-resolution Blackadar (1979) planetary boundary layer scheme (Zhang and Anthes 1982), and shallow cumulus parameterization (Grell et al. 1994). The parameterization choices are based on the work of Vizy and Cook (2002, 2003) in their simulations of Africa and the tropical Atlantic.

In the three climate-mode simulations, changes in soil moisture availability are used to manipulate the strength of the African easterly jet and then examine changes in the waves. Soil moisture availability, a dimensionless fraction, is used in the model as a multiplier in the surface moisture flux equation. Empirical conversion formulas in the soil bucket model are used to convert the observed or the NCEP soil moisture (in units of millimeters) to soil moisture availability, though no soil model is used in the present study.

Therefore, moisture availability is not exactly equivalent to the degree of saturation on the ground, but it is related to it. Effects of vertical soil moisture distributions are not included. One simulation has realistic soil moisture availability, which is updated every 12 h by linear interpolation based on the monthly soil moisture climatology of Willmott et al. (1985). Across northern Africa, the soil moisture availability ranges from about 0.6 over the West African rain forests to 0.02 over the Sahara, as shown in Fig. 2a. This range motivates two idealized soil moisture experiments, in which the soil moisture availability is uniform at values of 0.02 and 0.06, respectively. A land surface model is not used, and the soil moisture is prescribed and indifferent to the state of the atmosphere.

For the two synoptic-mode simulations, 120-km horizontal grid spacing is used on an even larger domain than in the climate-mode simulations, covering 14°S to 51°N in latitude and 125°W to 70°E in longitude (not shown). The same 24 vertical \( \sigma \) levels are used. The physical parameterizations for the synoptic experiments are those used in the climate-mode simulations, except the CCM2 scheme is used to parameterize radiation (see below) and the Grell scheme for cumulus convection.

Lateral boundary conditions for wind, temperature, and moisture are taken from the 1988 NCEP reanalysis. The 12-hourly time series is linearly interpolated to generate boundary conditions for each model time step of 3 min. Both of the 124-day synoptic simulations are initialized at 0000 UTC on 15 May using the May 1988 NCEP reanalysis conditions and run through the middle of September 1988. As in the climate-mode integrations, the first 17 days are discarded for model spinup.

Druyan and Fulakeza (2000) and Druyan et al. (2001) have also used a mesoscale model for synoptic simulations of African easterly waves, but on a smaller domain. They find that the simulated waves are sensitive to the treatment of initial moisture field and soil moisture in the model. In the synoptic simulations presented here, we needed 12-hourly soil moisture values.
for that particular year (1988). The only available “data” are from the NCEP reanalysis, even though these values are largely unvalidated and derived solely from the model fields forced by the data assimilation. Since the 1988 NCEP soil moisture on distribution does not look very realistic, as shown in Fig. 2b, another synoptic-mode simulation using observed soil moisture climatology (Willmott et al. 1985) was carried out. The small meridional gradient of soil moisture in the 1988 NCEP values produce a simulation with no jet, consistent with the results of Cook (1999).

Two modifications are made to compensate for the strong radiation interaction with low clouds that occurs when the CCM2 scheme is used. (The need for these modifications motivated our change to the RRTM radiation parameterization for the climate-mode simulations.) The first, similar to Giorgi et al. (1993), Vizy and Cook (2002), and Cook et al. (2003), is to specify clouds below 850 hPa transparent to incoming solar radiation. The second modification is to remove excessive moisture from the upper troposphere by fixing mixing ratio values in the top six σ levels (above 300 hPa) to their initial conditions, as in Vizy and Cook (2002) and similar to Horel et al. (1994). (The need for these modifications disappears with the use of the RRTM scheme.)

4. Model validation

Since the focus of this paper is the relationship between African easterly waves and a feature of the background climatology, namely, the African easterly jet, the model validation begins with an examination of the basic state in which the waves occur.

Latitude–pressure distributions of the July–August average zonal wind and potential temperature from the climate-mode simulation with realistic soil moisture and the NCEP reanalysis are displayed in Figs. 3a and 3b, respectively. Both are averaged from 20°E to 10°W longitude. Although reanalyses cannot be treated as pure observations, some fields like wind velocity and air temperature used for validation are strongly constrained by observed data (Kalnay et al. 1996), and they provide a validation opportunity in data-sparse regions such as West Africa. To compare with the NCEP reanalysis, the model variables are interpolated onto the coarser grid of the NCEP reanalysis, which is 2.5° by 2.5° for the horizontal resolution and has 17 vertical levels.

A more consistent comparison with the NCEP climatology would result from an ensemble average of the simulations for each year, but that is not practical. The magnitude of the simulated African easterly jet is 14 m s⁻¹, somewhat stronger than the 10 m s⁻¹ jet in the NCEP reanalysis, but well placed at 600 hPa and 13°N. The low-level westerly monsoon flow below the jet is also stronger in the model, so vertical and meridional wind shears are larger. The average magnitude of the jet in the climatology of the 1983–92 European Centre for Medium-Range Weather Forecasts (ECMWF) reanalysis is 8 m s⁻¹, and the low-level westerly flow is only 2 m s⁻¹, as shown in Fig. 3c. Some of the differences between the ECMWF and NCEP reanalyses may be related to the different averaging periods. The simulated isentropes are quite similar to those in the reanalyses. All show large deflections toward the ground north of 15°N, suggesting strong baroclinicity in association with strong sensible heating over the Sahara.

Figures 3d and 3e show the July–August mean zonal wind and potential temperature from the 1988 NCEP reanalysis and the synoptic-mode simulation with Willmott et al. (1985) soil moisture distributions, respectively. The 1988 jet is weaker than the climatological jet (Fig. 3b). (The 1988 jet in the ECMWF climatology is similar to that in the NCEP reanalysis, but the low-level westerly flow is weaker.) The 1988 simulated African easterly jet is broader and stronger than in the reanalyses.

Figure 3f displays the 1988 synoptic-mode simulation with NCEP soil moisture prescribed. The African easterly jet and the low-level westerlies are absent in this simulation. These differences are related to the use of a different soil moisture distribution, since the soil moisture’s control on the surface temperature is an impor-
tant factor determining the jet location and strength (Cook 1999).

The model produces realistic summer rainfall distributions over Africa and the tropical Atlantic. Vizy and Cook (2002) compared model results with various observational and satellite-gauge estimates and concluded that the model captures the summer rainfall over this region. Figure 4 displays the zonal mean power spectral den-
sity averaged from 10°W to 20°E from the realistic climate-mode simulation, computed from the time series of the meridional wind at 725 hPa from mid-June to mid-September. The maximum at a period of about 4.5 days between 10° and 15°N indicates the presence of 3–5-day African easterly waves, in agreement with the spectral analyses of previous studies (Burpee 1974; Diedhiou et al. 1998, 1999). North of about 25°N, there is strong wave activity between 4.5–7 days and 8–10 days, consistent with the 6–9-day African easterly waves identified in the reanalyses.

To further validate the modeled waves with the various easterly wave types distinguished by Viltard et al. (1997) and Diedhiou et al. (1998), bandpass filters are used to separate wave disturbances with 3–5-day periods from the 6–9-day waves. Figure 5 shows the response functions of the two bandpass filters used, which offer steep rolloff characteristics for the spectral windows. The two filtered time series are used to form composite waves as in Diedhiou et al. (1998) by selecting and averaging the 725-hPa filtered wind fields with filtered meridional wind speed magnitudes at 18°N and 0° greater than 1 m s⁻¹ (northward). The composite waves reflect an average picture of the significantly perturbed fields based on the reference point (18°N, 0°).

The 3–5-day composite waves are shown in Fig. 6a. They tilt from southeast to northwest north of the jet (between 10° and 15°N), and from northeast to southwest south of the jet, as noted in Diedhiou et al. (1998). The wave trains merge over the Atlantic, similar to Reed et al.’s (1988) findings from the ECMWF reanalysis. The axis of the cyclonic circulation at 25°N between 10° and 0° tilts from the southwest to the northeast, leading to negative barotropic conversions \( C_k = -\Delta \rho / g [u' v' \partial v'/\partial y] \) since \( [u' v'/\partial v'/\partial y] > 0 \) in westerly shears, that is, \( \partial v'/\partial y > 0 \) (\( u \) is zonal wind, \( v \) is the meridional wind, and \( \rho \) is pressure). In contrast to the positive barotropic conversions near the two sides of the jet, this cyclonic circulation cannot be obtaining eddy kinetic energy from the African easterly jet.

The vertical structure of the composite 3–5-day wave (not shown) is also in agreement with previous investigations (Burpee 1974, 1975; Reed et al. 1977). The wave trough axis tilts eastward with increasing height below the jet, and westward above the jet. This structure implies that energy is being converted from zonal mean to eddy available potential energy.

Figure 6b displays 6–9-day composite waves. These wave patterns are located farther north, in agreement with observations. Diedhiou et al. (1998) suggested that the 6–9-day waves originate from the Libyan and the Azores areas, and this is consistent with the modeled waves since the first anticyclonic circulation is centered around 25°N, 20°E (Libyan area), and the first cyclonic circulation is centered farther north around 35°N, 30°W (near the Azores). The 6–9-day wind perturbations are weak south of 5°N and stronger between 10° and 15°N.

Figure 7a shows Hovmöller correlation diagrams of the 3–5-day filtered meridional wind at 725 hPa, computed by correlating the time series of the filtered meridional wind at 15°N, 0° with that of each grid point along 15°N in different time lags (Liebmann and Hendon 1990). The slope of the contours indicates the westward propagation of the African easterly waves. At 15°N, the wavelength increases from 2400 km at 20°E to 3500 km at 10°W. The modeled wavelength also increases toward the equator, similar to the observational study by Diedhiou et al. (1999). The slope of the correlation isopleths in Fig. 7a gives the waves’ average
phase speed, which is about 8.3 m s\(^{-1}\) at 0°. At 15°N, it increases from 6 m s\(^{-1}\) at 20°E to 10 m s\(^{-1}\) at 10°W, similar to those of previous investigations (Reed et al. 1977; Diedhiou et al. 1998).

Figure 7b shows a Hovmöller correlation diagram for 6–9-day waves with the reference point at 22°N, 0°. The distortion of the correlation isopleths near 10°E is due to the Tahat Mountains (see Fig. 1). However, since the 6–9-day waves occur east of the Tahat Mountains, they are clearly not caused by the interaction of the flow with topography. The average wavelength is around 5300 km, the local phase speed is 6 m s\(^{-1}\). The wavelength and phase speed estimated by Diedhiou et al. (1998) are 5500 km and 6 m s\(^{-1}\), respectively.

5. Results

a. The climate-mode simulations

The relationship between the African easterly jet and the waves in the three climate-mode experiments is examined. As described in section 3, the “realistic case” has soil moisture availability prescribed based on observations, and two idealized cases have uniform soil moisture availability of 0.02 and 0.06, the “dry” and “wet” cases, respectively.

Figures 8a–c show horizontal wind vectors and contours of the zonal wind at 625 hPa, averaged from 15 June to 14 September for the realistic, dry, and wet cases, respectively. In the realistic and dry cases, the
African easterly jet extends to about 30°E and has maximum wind speeds of 15 m s\(^{-1}\) near the west coast, similar to the observations. The jet is a little weaker over West Africa in the dry case than in the realistic case in association with a slight decrease of the meridional surface temperature gradient over West Africa.

The jet is quite weak in the wet case, shown in Fig. 8c, with a maximum zonal wind speed of 9 m s\(^{-1}\) over Central Africa. This jet structure is associated with a decrease in meridional geopotential height gradients over West Africa. Strong evaporative cooling over the Sahara due to the unrealistically wetland surface cools this region and reduces the magnitude of the (positive) surface and, thereby, lower-tropospheric temperature gradient.

Composite 3–5-day waves at 725 hPa for the dry case, computed using 57 snapshots, are shown in Fig. 9a. The same reference point at 18°N, 0°, as in the realistic case, is used for the composite waves of both the dry and wet cases. The 3–5-day wave disturbances are active off the west coast of Africa, similar to the realistic case, but there is no robust wave structure over the African continent except, perhaps, within the ITCZ near 5°N. Thus, the presence of a realistic African easterly jet in this simulation (Fig. 8b) and, with it, the possibility of barotropic energy conversions to initiate African waves over the land surface, does not produce significant wave activity.

Figure 9b displays composite 3–5-day waves for the wet case. Despite the absence of a well-defined jet in the west, the waves are quite active. The fact that 204 snapshots exceed the 1 m s\(^{-1}\) threshold velocity also indicates that the wave activity is stronger than in the dry case. A realistic tilt of the wave axes is present (cf. with Fig. 6a), and the phase speed and wavelength of the wave disturbances are around 10 m s\(^{-1}\) and 3700 km, respectively, within the ranges reported in previous investigations (Burpee 1974; Diedhiou et al. 2001).

Figures 10a–c display the seasonal mean power spectral density averaged between 10° and 15°N of the 725-hPa meridional wind for the realistic, dry, and wet cases, respectively. (The coast is near 17°W.) In the realistic case, 3–5-day wave activity occurs as far east as 20°E, with a local maximum between 4 and 5 days near 7°E, and modes with periods between 3 and 4 days appear near the Greenwich meridian and over the eastern Atlantic. The 3–5-day wave activity is confined west of the Greenwich meridian in the dry case (Fig. 10b), consistent with the composite analysis (Fig. 9a), with strong activity centered at 11°W. The wet case, without a strong jet over West Africa, has the strongest wave activity, centered at about 2°E and a period of 4.5 days.

Strong waves may weaken the jet temporally through barotropic conversions, as noted by the previous simple modeling results. However, as long as the strong surface temperature gradient exists, the jet will restore its strength through a thermally direct ageostrophic meridional circulation (Cook 1999). Therefore, the weak jet in the wet case is not the result of strong waves but of a weaker surface temperature gradient.

As discussed in section 2, the Charney–Stern and Fjörtoft necessary conditions for instability have been associated with the generation of African easterly waves. To evaluate this association in the climate-mode simulations, Ertel IPV is computed. As given in Molinari et al. (1997),

\[
\text{IPV} = -g \frac{\partial \theta}{\partial p} (s_0 + f) = g \sigma^{-1} (s_0 + f),
\]

where \(\sigma = -\frac{\partial \rho}{\partial \theta}\) is the mass density, \(\theta\) is potential temperature, \(\zeta_0\) is the vertical component of relative...
Fig. 8. Seasonal mean zonal wind contours and wind vectors at 625 hPa from the climate model simulations with (a) realistic soil moisture, (b) a uniformly dry surface, and (c) a uniformly wet surface. The contour interval is 3 m s$^{-1}$, and the vector scale is indicated in m s$^{-1}$. Dashed contours represent easterly zonal flow, and solid contours represent westerly zonal flow.
vorticity computed on isentropic surfaces, $g$ is gravitational acceleration, $p$ is pressure, and $f = 2\Omega \sin \theta$ is the Coriolis parameter. Equation (1) indicates that the IPV anomaly is associated with both an absolute vorticity anomaly and with a mass density anomaly. Note that they are not independent. The former is related to barotropic instability, and the latter to baroclinic instability.

Figure 11a shows contours of the seasonal mean potential vorticity (solid line) and pressure distribution (dashed–dotted lines) averaged between $10^\circ$ and $25^\circ$E on isentropic surfaces for the realistic climate-mode simulation. This averaging region is chosen to emphasize the wave generation region and to explore the idea that the initial or early wave growth associated with the IPV gradient reversals is related to the jet’s wind shears. Over West Africa, strong convection effect may dominate over the effect of wind shears. The blank area denotes regions in which the isentropes dip under ground in the Sahara (Fig. 3a) some time during the summer analysis period. IPV is at a maximum at $9^\circ$N near the 320-K isentrope on the southern side of the African easterly jet, which is located near the 318-K isentrope between $10^\circ$ and $15^\circ$N (Fig. 3a). The meridional IPV gradient is larger on the southern side of this IPV anomaly than on the northern side due to the larger planetary vorticity gradient ($df/d\theta$) on the southern side. A second IPV maximum appears in the upper troposphere near the tropical easterly jet, which is located around $7^\circ$N near the 345-K isentrope, as noted in previous studies (Diedhiou et al. 2002).
Shading in Fig. 11a indicates areas with negative IPV gradients. In the region of negative meridional gradient of IPV between 9° and 15°N, the Fjørtoft necessary condition is also satisfied since the flow is easterly (see Fig. 3a). Note that this negative IPV gradient region extends from the lower troposphere to the tropopause, where the IPV increases dramatically as a result of the smallness of mass density, $\sigma$. This structure suggests that the horizontal and vertical shears of the mean zonal wind (i.e., the African easterly jet) are not the primary cause of the negative IPV gradients because these shears are confined below the 500-hPa (325 K) level (Fig. 3a). Thorncroft and Hoskins (1994a) reexamined the analytically defined jet flow (centered at 15°N) by Simmons (1977) and found that the negative IPV gradient is located between 10° and 20°N below 400 hPa. The negative IPV gradient north of 12°N below 600 hPa (not clearly shown in Fig. 11a because of the deflection of the isentropes over the Sahara) results from the IPV minimum north of 12°N in the lower troposphere in association with dry convection over
the Sahara, as noted by Thorncroft and Blackburn (1999).

Figures 11b and 11c show cross sections of seasonal mean streamlines and diabatic heating, respectively, averaged between $10^\circ$ and $25^\circ$E. The zonal mean confluent streamlines tilt southward with increasing height to $7^\circ$N as the warm dry air from the Sahara is lifted over cooler humid air from the south. Convection in the ITCZ reaches up to the 340-K isentrope (about 300 hPa), with two condensational heating maxima that are approximately proportional to the vertical $p$ velocity $\omega$ (not shown). The modeled precipitation maximum near $6^\circ$N is consistent with the observation (Nieuwolt 1977) that most of the rainfall within the ITCZ over Africa falls to the south of the surface convergence zone ($18^\circ$N; not shown in isentropic coordinate) since the uplifting air condenses near and above the midtroposphere. Air north of the confluence zone tends to be warm and dry. Because of the smaller inertial stability (smaller $f$) and larger moisture supply on the equatorward side of the ITCZ, there is stronger condensational heating.

Schubert et al. (1991) found that reversed IPV gradients can occur when concentrated regions of convection produce IPV maxima. Such a reversal would be produced on the poleward side of the ITCZ at low levels and on the equatorward side of the ITCZ at upper levels. In the simulation, a positive IPV anomaly occurs to the north of the convergent center near $7^\circ$N (Fig. 11a). A detailed analysis of the IPV budget (Hsieh and Cook 2004a, manuscript submitted to J. Atmos. Sci.) shows that this IPV anomaly is a result of IPV stretching and tilting caused by ITCZ convection, which suggests that the negative IPV gradient near the midtroposphere is primarily caused by ITCZ convection. IPV compression, associated with divergence near 345 K, induces anticyclonic vorticity and leads to the local minimum of IPV near the tropopause. These convective IPV sources in the midtroposphere (near 320 K) accompanied by sinking aloft produce the negative IPV gradient to the north of the ITCZ near middle levels ($9^\circ$ and $15^\circ$N) and to the south at upper levels ($4^\circ$ and $7^\circ$N).

The spatial and temporal scales of the convective system are close to the scales of synoptic-scale wave disturbances over West Africa because of the interactions between the convective-scale and the large-scale flow. For the seasonal mean climatology, synoptic-scale convective systems along the ITCZ can produce an IPV anomaly with a magnitude of about $3 \times 10^{-7}$ K m$^2$ kg$^{-1}$ s$^{-1}$ south of the jet (see Fig. 11a).

Figure 12 is the same as Fig. 11, but for the dry case. The magnitude and extent of the IPV maxima at $9^\circ$N in the midtroposphere are both smaller than in the realistic case, suggesting a smaller unstable region in the vicinity of the jet (centered at $12^\circ$N near the 318-K isentrope). The strength of the African easterly jet is comparable to that of the realistic case, which indicates that comparable zonal mean kinetic and available potential energy is present. However, there is no significant 3–5-day wave activity over Africa despite the IPV gradient reversal. Dickinson and Molinari (2000) obtained a similar result over northern Australia. They found little evidence for 2–10-day wave disturbances in the region and suggested that this may be because the unstable region is not long enough to support waves.

According to Figs. 12b and 12c, convection between $6^\circ$ and $9^\circ$N in the dry case is much more shallow and weaker than in the realistic case (Figs. 11b and 11c). Figure 12c displays the diabatic heating distribution for
the dry case. This result is similar to that of the observational analysis of Taylor et al. (2003), who suggest that soil moisture plays an important role in determining the stability of the atmosphere. The weakening of the ITCZ leads to a weaker IPV anomaly south of the African easterly jet, causing a smaller region of IPV gradient reversal between 9° and 13°N than in the realistic case.

Figure 13 is the same format as Fig. 11, but for the wet case. The IPV maximum near 12°N is comparable in extent to that in the realistic case, but slightly weaker and located about 2° of latitude farther north. As in the realistic case, IPV gradient reversals occur on the poleward side of the convergence maximum at lower levels near 9°N, and on the equatorward side aloft between 5° and 9°N. The total condensational heating is somewhat larger than that in the realistic case and nearly double the magnitude in the dry case, consistent with the presence of strong wave activity.

These simulations suggest that wave activity is associated more closely with intense convection than with the African easterly jet. Additional evidence comes from the wet case, in which wave activity becomes weak off the west coast of Africa in association with a rapid decay of convection (not shown). On the other hand, stronger convection off the west coast and over the tropical Atlantic in the dry case leads to more significant wave activity accompanying reversed the IPV gradients over the tropical Atlantic (not shown).

b. Results of the synoptic-mode simulations

In the synoptic simulation with NCEP soil moisture values (see section 3), the African easterly jet is absent and is replaced by weak westerly flow in the midtroposphere with weak easterly flow below 650 hPa (Fig. 3f). Figure 14a shows the mean zonal wind contours and wind vectors at 625 hPa for (a) the synoptic case with no jet using the 1988 NCEP soil moisture values, and (b) the synoptic case with the jet using the observed soil moisture climatology from Willmott et al. (1985). Dashed contours represent easterly zonal wind, and solid contours represent westerly zonal wind. Contour interval is 3 m s⁻¹, and the vector scale is indicated in m s⁻¹.
simulation with the jet (Fig. 15b) is obtained by averaging 206 snapshots, approximately twice the number as in the synoptic case without the jet, which suggests that waves are more active in the synoptic case with the jet than in the simulation without the jet. Wavelengths and phase speeds are more realistic, averaging 3300 km and 8 m s\(^{-1}\), respectively. Also, the wave axes tilt north and south of the jet (centered near 11\(^\circ\)N), as in the realistic climate-mode simulation and the observations. Strong wave amplitudes over West Africa are associated with the stronger ITCZ. The jet strength does not increase much over West Africa, as shown in Fig. 14b.

An examination of power spectral density distributions (not shown) for the two synoptic simulations supports the results from the composite view of the waves in Figs. 15. Without the jet, wave activity is present, but it is relatively weak and does not change much with longitude. Wave activity in the synoptic case with the jet (Fig. 15b) is strong over West Africa and has both realistic periods and wavelengths.

As reviewed above, simple model applications suggest that African easterly waves are caused by barotropic instability associated with the jet (e.g., Simmons 1977; Kwon 1989). Some observations, on the other hand, indicate that baroclinic conversions are more important over land (e.g., Norquist et al. 1977). The viewpoints can be reconciled if the wave initiation is by barotropic conversion, with the wave initiation perhaps missed by our current observing systems, with baroclinic processes becoming dominant as the waves propagate to the west (e.g., Thorncroft and Hoskins 1994a,b). Our results, however, do not support this idea. In the simulation without a jet and, therefore, no barotropic conversions, waves are certainly initiated unrealistically wet conditions over the Sahara for 1988. We use this simulation as an idealization, to understand whether or not wave activity can be generated in the absence of the jet. It is compared with the synoptic simulation with Willmott et al. (1985) climatological soil moisture values, which produces a reasonable African easterly jet at 625 hPa, as shown in Fig. 14b.

Figure 15a displays composite meridional wind contours of 3–5-day wave disturbances at 725 hPa for the synoptic case without the jet, using 12\(^\circ\)N, 0\(^\circ\) as the reference point. Despite the lack of an African easterly jet, easterly waves are still generated, but the 1100-km average wavelength is smaller than the reported values of the previous studies (2000–4000 km). The phase speed is approximately 6 m s\(^{-1}\) westward. Since the absence of the jet eliminates barotropic conversions and the instability from the strong surface heating in the simulation, the waves are being maintained by baroclinic energy conversions due to condensational heating. Some wave disturbances may be triggered by finite-amplitude perturbations in a neutral basic state. However, without the condensational heating source, these finite-amplitude disturbances in a neutral basic state would eventually dissipate due to friction since they cannot obtain energy from a neutral zonal flow.

The shorter wavelength of the waves embedded in a weak zonal flow can be explained by considering a pressure perturbation due to a convective heating pulse, causing the flow to deviate into the state of gradient motion, which is governed by

\[
\frac{v}{r} = -f + \frac{1}{v} \frac{\partial \phi}{\partial r},
\]

where \(v\) is wind speed, \(\phi\) is geopotential, and \(r\) is radius of curvature. The gradient wind speed \(v\) can be expressed as follows:

\[
v = -f r \pm \frac{\sqrt{r^2 + r^2}}{4 + r^2} \frac{\partial \phi}{\partial r}.
\]

If the heating is very near the equator (\(f \approx 0\)), then \(r\) is approximately proportional to \(u\). If \(f\) is relatively large (farther off the equator), for example, with \(f\) and \(1/v\) \((\partial \phi/\partial r)\) of similar size, then \(r\) is proportional to \(v\). Therefore, the size of the generated vortex is approximately proportional to the scale between \(v\) and \(\partial \phi/\partial r\), depending on the relative magnitudes of \(f\) and \(1/v\) \((\partial \phi/\partial r)\). Thus, a smaller-scale wave could be expected in weaker zonal flow. This suggests that the jet plays a role in setting the scale of the waves, but the analysis shows that it is not directly involved in their generation.

The smaller wavelength response to weaker surface temperature gradient was also noted by Chang (1993), who found that below the near-neutral layers over the Sahara the most unstable range shifts toward shorter waves as meridional temperature gradient decrease.

The composite meridional wind pattern computed with the reference point at 12\(^\circ\)N, 0\(^\circ\) in the synoptic
over central Africa (Fig. 15a), and they are as strongly represented in this region as in the case with a realistic jet.

Figures 16a and 16b display the seasonal mean IPV and pressure distributions on isentropes averaged between 10° and 25°E for (a) the synoptic case without the jet and (b) the synoptic case with the jet. Contour interval is 10 \(^7\) K m\(^2\) kg\(^{-1}\) s\(^{-1}\). Shading indicates regions with negative meridional IPV gradient.

In the synoptic simulation with the jet (Fig. 16b), there is a region of reversed IPV gradient near 320 K between 9° and 15°N that coincides with the easterly jet, so both the Charney–Stern and the Fjörtoft necessary conditions are satisfied in this region. The weaker IPV anomaly is due to the weaker ITCZ convection over central and East Africa, as noted in section 3. The mean variance maximum of meridional wind perturbations (not shown) is located between 10° and 15°N near 315 K, which coincides with the unstable region suggested by the IPV gradient reversal.

The 3–5-day easterly waves are generated in both synoptic simulations, despite the fact that one does not have an African easterly jet and the Charney–Stern and Fjörtoft instability criteria are not met. In this no-jet case, convective heating alone generates the waves. But the waves are more realistic in the simulation with a jet, especially to the west of the initiation region.

6. Summary and conclusions

Regional model simulations are analyzed to explore the relationship between African easterly waves and the basic-state background climatology. The individual relationships of the waves to the African easterly jet and the ITCZ are emphasized, though they are not two completely independent climatological features over Africa, as noted by Thorncroft and Blackburn (1999). The model is run in both climate and synoptic mode, with various prescriptions of surface moisture used to modify, and even eliminate, the African easterly jet. In climate mode, surface and lateral boundary conditions are taken from climatologies. The synoptic-mode simulations are aimed at capturing a particular time and use boundary conditions for the summer of 1988.

More insight into the wave-generating mechanisms is sought by considering how the background atmospheric state generates the meridional IPV gradient reversals required by the Charney–Stern necessary (but not sufficient) instability criterion. The two possibilities are strong wind shears near the jet, and very concentrated condensational heating aloft (in the ITCZ). The former is associated with combined barotropic/baroclinic energy conversions from the jet, and the latter with induced barotropic and baroclinic conversions due to convection. (Cumulus convection within the ITCZ not only induces baroclinic conversions due to vertical motion, but also barotropic conversions as cyclonic relative vorticity is strengthened through vortex stretching.) The positive IPV anomaly (IPV maximum) at 9°N near the midtroposphere primarily results from moist convection south of the African easterly jet, while the IPV anomaly (IPV minimum) north of the jet in the lower troposphere is caused by dry convection over the Sahara.

In the climate-mode experiments, sign reversals of IPV gradients occur in all three cases, even though no significant wave activity occurs over Africa in the dry case. Apparently, the reversal of the IPV gradient does not necessarily lead to an unstable zonal flow since it is not a sufficient condition. Cycloic IPV anomalies,
caused by vortex stretching resulting from cumulus convection, reverse the meridional IPV gradient on the poleward side of the ITCZ, where the zonal flows are expected to be unstable.

Barotropic instability of the jet does not appear to be the key factor for the initiation of the waves in the climate-mode simulations, since no significant waves are generated over Africa in the dry case with a strong jet. In contrast, stronger waves occur over Africa in the wet case without a strong easterly flow. In the synoptic simulation with no jet, the Charney–Stern instability criterion is not met in the vicinity of the 3–5-day wave disturbances, again suggesting that that these waves can be generated in the absence of shear instabilities. But without the presence of the African easterly jet, the scale of the wave disturbances is smaller due, perhaps, to a lack of interaction with the large-scale zonal flow.

Overall, these five simulations suggest that the African easterly jet is of less importance than the convection associated with the ITCZ as a cause of wave activity. A detailed analysis of the IPV budget, including associations with barotropic and baroclinic instability, is currently being carried out to better understand how IPV gradient reversals are associated with the generation of the waves.

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