Energy Dispersion in African Easterly Waves

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ABSTRACT

The existence of an upstream (eastward) group velocity for African easterly waves (AEWs) is shown based on single-point lag regressions using gridded reanalysis data from 1990 to 2010. The eastward energy dispersion is consistent with the direction of ageostrophic geopotential flux vectors. A local eddy kinetic energy (EKE) budget reveals that, early in the life cycle of AEWs, growth rate due to geopotential flux convergence exceeds baroclinic and barotropic growth rates. Later in the life cycle, EKE decay due to geopotential flux divergence cancels or exceeds baroclinic and barotropic growth. A potential vorticity (PV) budget is used to diagnose tendencies related to group propagation. Although both upstream and downstream group speeds are possible because of the reversal in the mean meridional PV gradient, upstream propagation associated with the positive poleward PV gradient dominates wave packet evolution. Analogous to the concept of downstream development of midlatitude baroclinic waves, new AEWs develop preferentially upstream of the older ones, thus providing a mechanism for seeding new waves. It is suggested that these results are also relevant to AEW intermittency and storm-track structure.

1. Introduction

a. Background

African easterly waves (AEWs) are synoptic-scale disturbances of importance to the summer weather over North Africa. They are also precursors to the majority of Atlantic tropical cyclones (e.g., Pasch et al. 1998). Hydrodynamic instability of the midtropospheric African easterly jet (AEJ) is the traditional starting point to understand the dynamics of AEWs. Burpee (1972) showed that the AEJ is associated with a reversal in the meridional potential vorticity (PV) gradient. This reversal satisfies the necessary condition for barotropic instability (Charney and Stern 1962) and is caused by the near-zero PV associated with shallow, dry convection to the north within the Sahara, and larger PV values associated with deep, moist convection within the intertropical convergence zone to the south (Thorncroft and Blackburn 1999).

In a statistical sense, the observed three-dimensional structure of these waves can be modeled by the fastest-growing normal mode of the climatological basic state over North Africa (e.g., Thorncroft and Hoskins 1994; Hall et al. 2006). Energy budgets calculated using observations and model simulations show that both barotropic and baroclinic energy conversions contribute to AEW development (e.g., Norquist et al. 1977; Thorncroft and Hoskins 1994; Hall et al. 2006). Many studies have also identified two preferred AEW tracks (e.g., Pytharoulis and Thorncroft 1999; Chen 2006). The one north of the AEJ is dominated by dry, shallow systems growing along the surface baroclinic zone near 20°N, and the one south of the AEJ is dominated by moist systems growing at jet level. The growth of the former is primarily through baroclinic conversion and that of the latter is through barotropic conversion (Kiladis et al. 2006).

b. AEW initiation, amplification, and intermittency

Although barotropic and baroclinic energy conversions are useful for describing the amplification of AEWs, a growing body of research suggests that normal-mode instability of the AEJ may be inadequate to explain fully the observed life cycle of AEWs. Hall et al. (2006) demonstrated that a relatively small amount of low-level damping in a numerical simulation can stabilize the AEJ to small perturbations. Using the same model configuration, Thorncroft et al. (2008) found that AEWs can be generated if a large region of diabatic heating that
mimics the effect of convection is applied near the entrance region of a linearly stable AEJ. This finding supports the hypothesis that AEW initiation requires a large-amplitude precursor. Many studies have linked individual AEWs with convective precursors over eastern Africa (e.g., Carlson 1969; Berry and Thornycroft 2005; Mekonnen et al. 2006; Hsieh and Cook 2008). Hsieh and Cook (2005) found that individual convective events themselves may lead to the PV gradient reversal that initiates AEWs.

As the mechanisms for AEW growth and initiation are becoming better understood, recent work has focused on their intraseasonal and interseasonal intermittency. In other words, what causes periods of increased or decreased AEW activity within and between seasons? Using a dry numerical model with AEWs triggered by localized heating, Leroux and Hall (2009) found that AEWs develop most readily when the AEJ has a strong PV gradient over a large region. However, in a similar numerical model, Hall et al. (2006) compared the most unstable normal modes of a basic state with abundant AEW activity (July-August 1988) and one with little AEW activity (July-August 1990), and they found little difference in the simulated wave activity. The mean state of the AEJ was nearly similar in both seasons, and the simulated normal modes were also nearly identical. This is in contrast with marked differences in the observed wave activity during these two seasons. This suggests that the normal-mode instability approach is inadequate to explain AEW intermittency and variability.

Based on the difficulty of linking AEW intermittency with the state of the AEJ, many studies have suggested that the variability of convection is a more important factor controlling periods of active and inactive AEWs than is the state of the AEJ. Using a regional climate model, Hsieh and Cook (2005) found that AEWs develop more readily in a simulation with a weak AEJ and a strong ITCZ than they do with a strong AEJ and a weak ITCZ. They even suggest that the AEJ may be unnecessary in the triggering of AEWs, since they form in simulations without an AEJ. Cornforth et al. (2009) point toward an internal mechanism that involves the interaction of moist processes with the AEJ and AEWs for setting wave intermittency. Mekonnen et al. (2006) found that most AEWs can be tracked backed to convective triggers in the highlands of East Africa. They also attribute the marked differences in AEW activity between 1988 and 1990 to 1988 being a much wetter year. However, it is unclear from their study to what extent the stronger AEW activity contributes to the increased rainfall. Leroux et al. (2010) also find observational evidence that increased moist convection in East Africa is often followed by an increase in AEW activity. Leroux et al. (2011) found using an idealized numerical model that extratropical forcing from the North Atlantic storm track is at least as effective at generating AEWs as local convection, and Ventrice et al. (2011) connect AEW activity with the Madden-Julian oscillation. Since none of these mechanisms is mutually exclusive, it is likely that a combination of all of these factors leads to large intraseasonal variability of AEWs.

c. Energy dispersion and wave life cycle

It is apparent that normal-mode hydrodynamic instability does not adequately explain the life cycle of AEWs. In this context, it may be useful to draw an analogy with baroclinic waves in the midlatitudes. The traditional description of the midlatitude wave life cycle relies on normal-mode baroclinic instability of the background state (e.g., Eady 1949). However, alternative theory- and observations-based studies have suggested that they are more appropriately described as modulated wave packets evolving within a streamwise varying background (e.g., Chang 1993; Mak and Cai 1989). Therefore, group propagation and its attendant properties, such as dispersion of energy within the wave packet, become preeminent in determining the evolution of baroclinic waves (Chang et al. 2002).

The direction of energy propagation of midlatitude baroclinic waves is primarily eastward, and thus this process is commonly referred to as downstream development. The earliest theoretical studies of this phenomenon related it to the eastward group velocity of linear barotropic Rossby waves (Rossby 1949). Motivated by observations of downstream development of synoptic-scale baroclinic waves, Simmons and Hoskins (1979) examined this phenomenon for unstable baroclinic Rossby waves using a dry, idealized numerical model. They found that new disturbances form downstream of old ones at upper levels and upstream at lower levels. They conclude that the main trigger for extratropical cyclones may not be small-amplitude perturbations growing through baroclinic instability but rather energy dispersion from neighboring disturbances. Orlanski and Katzfey (1991) demonstrated this process for a single case in the South Pacific, and Chang (1993) showed it using statistical regression of data along the North Pacific storm track.

The wave packet description of the midlatitude waves has provided important insights not readily accessible through the modal instability paradigm. It is consistent with observations that large-amplitude precursors are readily available for seeding new disturbances. It also provides a better model for the observed length and structure of the midlatitude storm track (Chang et al.
In particular, downstream development allows for an extension of the storm track into regions of weaker baroclinicity. It is of interest to determine whether a similar approach will also aid in improved understanding of the AEW life cycle and storm track. To the extent which AEWs are analogous to midlatitude baroclinic waves, they should exhibit similar energy dispersion and group velocity characteristics. However, unlike baroclinic waves in the midlatitudes, AEWs develop on an easterly jet and grow through both baroclinic and barotropic energy conversion. Additionally, the relative contributions of energy through diabatic heating may be more significant for AEWs than that of midlatitude cyclones, and the stronger gradient of planetary vorticity in lower latitudes implies stronger energy dispersion and faster group velocity. Nevertheless, it is anticipated that the concept of energy dispersion may be useful in describing the initiation, amplification, and decay of AEWs.

d. Objectives

The goal of this study is to examine whether observed AEW wave packets have distinct eastward group propagation. This is diagnosed using Hovmöller plots of meridional wind, an eddy kinetic energy (EKE) budget, and a PV budget. Finally, some implications of the upstream group propagation for the life cycle and storm track of AEWs are discussed.

2. Methods

a. Data

The gridded European Centre for Medium-Range Weather Forecasts (ECMWF) Interim Re-Analysis (ERA-Interim) (Dee et al. 2011) from 1990 to 2010 is used for the analysis. These data have a grid spacing of 1.5° × 1.5° and are available every 6 h. Rainfall rates from the Tropical Rainfall Measuring Mission (TRMM) 3B42 dataset are used to diagnose active convection. This dataset covers the period 1998–2010 and was re-gridded to the same resolution as the ERA-Interim for our analysis. Nicholson et al. (2003) found good agreement between TRMM-adjusted precipitation amounts and rain gauge data in West Africa.

b. Regression

One-point lag regressions are used to succinctly describe the time evolution of AEWs and their energetics. This technique has been applied to both extratropical cyclones (e.g., Chang 1993) and AEWs (e.g., Kiladis et al. 2006). The location (10°N, 15°W) near where AEWs reach their climatological maximum amplitude is chosen as a base point for regression. A broad, zonally oriented region of maximized PV flux divergence is centered at this latitude (Fig. 7 in Mekonnen et al. 2006). The results from longitudes between 40° and 0°W along 10°N are qualitatively similar. A reference time series at this base point is constructed using the 2–10-day bandpass-filtered 700-hPa meridional wind, which is one of the most trackable features of AEWs over a wide range of longitudes [see Fig. 6 in Mekonnen et al. (2006)]. The lower cutoff of 2 days removes the strong diurnal cycle of temperature and convection, and the upper cutoff of 10 days removes intraseasonal variability. This time filtering is somewhat longer than the 3–5-day or 2–6-day bands used in many other studies of AEWs. A wider range is important because a broad spectrum of wavelengths is needed to preserve the constructive and destructive wave interference necessary to observe wave packets with group propagation (e.g., Chang 1993). Relevant atmospheric fields are regressed against the reference time series. The standard deviation of the 700-hPa meridional wind at the base point is used to scale the regression coefficients. The result is a three-dimensional composite of an AEW from which an energy budget is calculated.

c. Energy budget

Following Orlanski and Katzfey (1991), an EKE budget is used to diagnose energy dispersion. This procedure involves partitioning the kinetic energy into a time-mean component and a deviation from it. Many previous studies have relied on volume and temporally averaged EKE budgets to characterize the energy exchange processes between AEWs and the AEJ (e.g., Norquist et al. 1977; Hsieh and Cook 2007). However, to understand the energy transfer of EKE between individual AEWs, a local energy budget is needed. Orlanski and Katzfey (1991) derived the following equation for the budget of EKE in isobaric coordinates:

\[
\frac{\partial (K_e)}{\partial t} + \mathbf{V}_m \cdot \nabla K_e + \mathbf{V} \cdot \nabla_3 K_e = - (\mathbf{V} \cdot \nabla \phi) - [\mathbf{V} \cdot (\nabla_3 \mathbf{V}_m)] + [\mathbf{V} \cdot (\nabla_3 \mathbf{V})] \\
- \text{diss}_e + \mathbf{F}_o ,
\]

where \(K_e\) is the EKE, \(\mathbf{V}_m\) is the time-averaged velocity, \(\mathbf{v}\) is the perturbation velocity, \(\phi\) is the perturbation geopotential, diss, refers to dissipative forcing, and \(\mathbf{F}_o\) denotes the forcing that maintains the time-averaged circulation. The terms on the lhs of Eq. (1) are the local EKE tendency, advection of EKE by the mean wind, and advection of EKE by the eddies, respectively. The first term on the rhs is the work done by the pressure field. The second and third terms denote the exchange of kinetic energy between the mean flow and the eddies,
respectively, commonly referred to as barotropic energy conversion. The fourth term is dissipation of kinetic energy by the eddies, and the last term is the impact of the steady-state forcing on the eddies and is generally found to be small.

Orlanski and Katzfey (1991) showed that the pressure work term is useful in diagnosing energy dispersion through its relationship with ageostrophic geopotential fluxes. This term, when vertically integrated, is approximated as

\[-v \cdot \nabla \phi = -\nabla \cdot (v \phi) - \omega \alpha, \tag{2}\]

where \(\omega\) is the vertical velocity in pressure coordinates and \(\alpha\) is the specific density. The second term on the rhs of Eq. (2) is the baroclinic term, which describes the conversion of eddy potential to eddy kinetic energy. Though it is often associated with horizontal temperature fluxes, it can also be associated with deep, moist convection. The first term on the rhs of Eq. (2) is the geopotential flux convergence (GFC). It is the focus of this study because of its relationship with energy dispersion and group velocity. Since it averages to zero over a global domain, previous energetics studies of AEWs have not addressed it. While the calculation of GFC is straightforward with reanalysis data, determining the direction of the flux is somewhat arbitrary. Following Orlanski and Katzfey (1991) and recognizing that geostrophic wind on the \(f\) plane is nondiagonal, the geopotential flux that yields nonzero flux convergence can be written as follows:

\[v_a \phi = \left( v - \frac{k}{f_0} \times \nabla \phi \right) \phi \tag{3}\]

where \(f_0\) is the Coriolis parameter at a reference latitude. Orlanski and Katzfey (1991) refer to the term on the lhs, which has the geostrophic component of the flux removed, as the ageostrophic geopotential flux. The same for a variable Coriolis parameter was shown by Orlanski and Sheldon (1993) to be

\[v_a \phi = v \phi - k \times \frac{\phi^2}{2f(y)} \tag{4}\]

While both of these equations work well in the mid-latitudes, applying either of them near the equator presents difficulties. If Eq. (3) is used, then the approximation of the ageostrophic wind retains an increasing large rotational (i.e., nondiagonal) component at latitudes far from the reference latitude. The larger gradient of \(f(y)\) near the equator magnifies this problem. If Eq. (4) is used instead, then the geostrophic wind equation has a singularity at the equator and thus becomes an increasingly inaccurate approximation for the actual wind. This inaccuracy yields an unrealistically large rotational component to the flux vectors in the region of interest, since it produces very large values of ageostrophic wind within a few degrees of the equator. Experimenting with both methods, we found Eq. (3) to be less deficient than Eq. (4) in its depiction of the ageostrophic geopotential flux vectors. However, because of the large gradient of \(f(y)\) near the equator, the choice of reference latitude is important. Since we are most interested in waves at low latitudes, we use a reference latitude of \(10^\circ\)N. Varying the reference latitude from \(5^\circ\) to \(20^\circ\)N does not alter the interpretation of our results.

d. Potential vorticity

The relationship between phase and group velocity for AEWs can also be diagnosed using the conservation equation for PV on an isentropic surface. Though it is difficult to quantify the growth of a single AEW solely due to group velocity effects using PV as compared to the local EKE approach, it nevertheless provides an intuitive illustration of wave packet evolution. Here, again, an analogy with barotropic Rossby wave packets is useful. From a PV perspective, the phase velocity will be directed toward where the tendency due to advection by a PV anomaly matches the sign of the anomaly itself, and the group velocity will be directed toward where the tendency is of the opposite sign (e.g., Hoskins et al. 1985; Nielsen-Gammon and Lefevre 1996). The group velocity of a wave packet results from the difference in strength of the anomalies associated with the individual waves within the packet. Partitioning the PV into time-mean and perturbation terms, the following form of the local tendency of perturbation PV is used:

\[\frac{\partial q'}{\partial t} = -v \cdot \nabla q' - v' \cdot \nabla q - v' \cdot \nabla q' + S, \tag{5}\]

where \(v\) is the velocity vector and \(q\) is PV on an isentropic surface. The terms with overbars represent the time-mean fields and those with the primes represent the perturbations. The first term on the right describes the advection of PV anomalies by the mean flow. Assuming the PV anomalies in the regressed fields are attributable to Rossby waves, the second term describes their phase and group velocity. The third term describes the nonlinear advection of the perturbation PV by perturbation velocity. The nonadvective sources of perturbation PV, which include diabatic effects and friction, are denoted by the last term (\(S\)) but are not diagnosed here. Because the diabatic generation of PV through convection is not available in the ERA-Interim, we use regressed TRMM
precipitation rates to identify regions of active convection where PV is likely being generated. Though this approach is not ideal, the impact of convection on phase and group velocity is beyond the scope of this study. For a discussion of the mean PV field over northern Africa, refer to Dickinson and Molinari (2000).

3. Results and analysis

Figure 1 shows a Hovmöller plot of 2–10-day filtered 850-hPa EKE averaged between 8° and 15°N for the summer of 2006. Westward-propagating AEWs are clearly seen in the sloping lines of EKE. Two additional features of AEW activity are also evident here.

- The intermittency of the waves as seen in periods of increased activity that are interspersed with periods of weaker activity. During this season, AEWs seem to be organized into three to four distinct wave packets.
- The tendency for EKE to be enhanced on the upstream (eastern) side the wave packets compared to the downstream (western) side. This is particularly clear in the AEW wave packet seen between 18 August and 1 September. This suggests that, in addition to a westward propagation of the wave phase, an eastward group propagation can be inferred.

Figure 2 depicts the composite evolution of wave packets using 850-hPa meridional wind regressed against the time series of 2–10-day filtered 850-hPa meridional wind at the 10°N, 15°W base point. While individual AEW troughs and ridges move westward, the maxima in meridional wind move slowly eastward. This pattern is consistent with westward phase and eastward group propagation. Unfiltered regressions provide a similar interpretation, though the signal is slightly weaker due to the strong diurnal cycle. Based on this plot, the zonal group velocity is eastward at 3.3 m s\(^{-1}\) and the zonal phase velocity is westward at 10.4 m s\(^{-1}\). However, since the AEWs are embedded within the westward flow of the AEJ, the flow relative group velocity is much faster toward the east. Even as they progress through the eastern Atlantic, AEWs continue to exhibit a signal of group velocity. Figure 3 shows a similar regression performed against a reference time series of 2–10-day filtered meridional wind at the 10°N, 30°W base point. A as in Fig. 2, the group velocity is eastward at 3.6 m s\(^{-1}\) and the phase velocity is westward at 10.1 ms\(^{-1}\). Following the meridional wind maxima, it can be seen that the 0.3 m s\(^{-1}\) signal extends 2.5 periods to lag = 7 days and only 1 period to lag = −3 days (i.e., in addition to the strong group velocity signal, it is also apparent that the meridional wind pattern is much more coherent into the lead times). This observation suggests that as AEWs move westward into the eastern Atlantic and decay, they continue to disperse their energy upstream toward Africa.
a. Kinetic energy budget

1) PLAN VIEW

To examine the mechanism by which EKE maxima associated with AEWs propagate upstream, the baroclinic, barotropic, and GFC components of the EKE budget are calculated. The advection terms on the lhs of Eq. (1) are not considered here because they merely act to translate the EKE centers westward. The individual terms of the EKE budget are computed using the perturbation fields obtained by regression against the time series of 2–10-day bandpass-filtered 700-hPa meridional wind at the 10°N, 30°W base point. The budget terms are scaled by the standard deviation of the meridional wind at the base point and vertically integrated from 950 to 350 hPa. The selection of the lower boundary of 950 hPa minimizes the amount of data used that is below the surface. The upper boundary is chosen because AEWs are concentrated in the lower half of the troposphere. The interpretation of the results is insensitive to these boundaries.

Figures 4–6 show the EKE budget terms at lag = 0, 24, and 48 h, respectively. The vertically integrated EKE is shown in black contours on each plot. The contour interval for EKE is in $\log_2$ starting at 0.2 J kg$^{-1}$. This nonlinear scale proved necessary to view the EKE centers over several days without an excessive number of contours. The vectors and shaded values on each plot are as follows: (a) 950-hPa perturbation winds and barotropic conversion [second term on the rhs of Eq. (1)], (b) 650-hPa perturbation winds and barotropic conversion [second term on the rhs of Eq. (2)], (c) vertically integrated ageostrophic geopotential flux vectors and geopotential flux convergence [first term on the rhs of Eq. (2)], and (d) full 650-hPa winds and the sum of the terms in (a)–(c). Figure 7 shows the residual for Eq. (1) at lag = 0 h. To diagnose regions of active convection, precipitation rates from TRMM are plotted in Fig. 8. To focus the discussion, we label three different EKE centers: one downstream over the eastern Atlantic (A), one near the base point (B), and one upstream to the east (C).

At a lag = 0 h, EKE centers A and B are well developed, while center C is much weaker (Fig. 4). For center A, the positive values of the barotropic and baroclinic terms contribute to EKE growth, while the geopotential flux divergence contributes to its decay. The net effect is a loss of EKE for center A as seen from its decay during subsequent times. Thus, despite extracting energy from the basic-state wind shear, most of the energy from center A is being fluxed upstream toward center B, as shown by the flux vectors in Fig. 4c.

Within center B, the baroclinic and barotropic terms contribute strongly to EKE generation. The baroclinic term is largest in the vicinity of the surface baroclinic zone at the southern edge of the Sahara around 18°N (Fig. 4b). In this zone, the southerly 950-hPa wind blowing across the poleward-oriented temperature gradient leads to cold advection and contributes to a southward heat flux. Thus, center B is associated with sinking cold air. The correspondence of the strongest 950-hPa wind perturbation in this zone is consistent with the understanding that low-level AEWs north of the AEJ are strongly driven by baroclinic growth (Pytharoulis and Thorncroft 1999). The barotropic term is largest to the north and south of the EKE maximum, where the 650-hPa perturbation wind tilts against the mean horizontal wind shear associated with the AEJ. Additionally, since moist convection tends to be located within the trough axis between centers A and B (Fig. 8), the overturning circulation associated with this convection may lead to both baroclinic and barotropic generation of EKE.

The GFC contributes to EKE growth on the downstream side of center B and to decay on its upstream side. The ageostrophic geopotential flux vectors suggest that center B is gaining EKE from center A and exporting it to center C. Interestingly, the strong baroclinic EKE generation in the north is largely canceled out by the strong geopotential flux divergence that exports EKE upstream to center C. This is consistent with Orlanski and Sheldon (1995), who noted that concentrated areas of $-\omega\alpha$ are often counteracted by the geopotential flux divergence. Overall, however, center B’s
growth is strongly positive, especially in its southern half, where all three terms are positive (Fig. 4d). Meanwhile, center C’s growth is dominated by GFC, with the fluxes emanating from center B. The contributions from baroclinic and barotropic growth are negligible at this stage. However, some baroclinic growth may result from its collocation with moist convection (Fig. 8b). Convection within the northerly flow of AEWs over West Africa to the east of 0 \degree W is consistent with the findings of Kiladis et al. (2006).

Figure 7 shows the residual Eq. (1). It includes both physical components, such as frictional dissipation and contributions from other time scales outside of the 2–10-day filter, and nonphysical components, such as discrepancies resulting from merging observational data with model output. The negative values in the northern region of center B likely result from frictional dissipation, since the 950-hPa winds are strong (Fig. 4a) and the Saharan boundary layer is very deep. Moving the limits of integration 100 hPa higher greatly reduces the magnitude of the residual in this region. Thus, it appears the generation of EKE is largely counteracted by dissipation of EKE through friction. The negative residual EKE tendencies in the southern portion of centers A and B is more difficult to explain. However, their magnitudes are smaller than the GFC term. Thus, their existence does not fundamentally alter the conclusion that GFC is important in the evolution of AEW packets. Nevertheless, it would be useful to perform a similar budget on a model simulation where all of the sources and sinks of EKE can be accounted for.

At lag = 24 h (Fig. 5), center A is much weaker as its EKE is being transferred to center B. Center B continues with relatively large baroclinic and barotropic growth. However, most of center B is now located within a region of negative GFC. As a result, center B is fluxing more energy to center C than it is receiving from center A. The combination of these three EKE budget terms leads to regions of net EKE decay on the eastern side of center B, with a small region of EKE growth on the western side. Farther upstream, center C continues to grow. Its growth is dominated by barotropic conversion and GFC from fluxes emanating from center B. According to Fig. 8c, convection is shifting into the southerly flow between centers A and B, consistent with Kiladis et al. (2006) for AEWs over the eastern
Atlantic. Over land, precipitation remains collocated with center C in the northerly flow.

At lag = 48 h (Fig. 6), center A has disappeared for the given contour interval. Meanwhile, center B continues with baroclinic and barotropic energy conversion though most of this energy is being exported upstream to the leading edge of center C. The combination of these factors results in an overall decay of EKE for center B. Center C is near its peak growth rate with positive contributions from all three terms. Upstream of center C, GFC is positive and eventually leads to the development of the next EKE center (not shown). The precipitation remains in a similar phase as before, in the southerly flow well offshore around 32°W, and in the northerly flow near the coast (Fig. 8d).

2) HOVMOELLER

To view the EKE budget terms succinctly over several AEW periods, Hovmöller plots of the individual EKE budget terms are presented. Because of the large latitudinal differences in the relative contributions from the individual terms, results are averaged over both 6°–14°N (Fig. 9) and 15°–20°N (Fig. 10). In the former, the barotropic term is strongest and cumulus convection tends to be more active. In the latter, the baroclinic term is strongest and cumulus convection is less active.

Figure 9 clearly reveals an eastward group velocity; the earliest EKE maximum is near 27°W and the latest EKE maximum is near 17°W. A coherent EKE signal extends slightly farther into the lag times than into the lead times. This pattern is consistent with weaker AEW activity east of the prime meridian and stronger activity over West Africa and the eastern Atlantic. The barotropic term (Fig. 9b) is well correlated with the EKE maxima at all longitudes, whereas the baroclinic term (Fig. 9a) appears less important for EKE growth. This result is consistent with the observation that southern track AEWs have a larger component of growth from barotropic conversion (Kiladis et al. 2006).

The GFC term (Fig. 9c) exhibits a wavelike pattern with a similar period and phase velocity as the EKE maxima. Early in their life cycle, EKE maxima are located mostly within regions of positive GFC, and later in their life cycle, they fall more within regions of negative GFC. Additionally, EKE maxima at lag times fall more within regions of negative GFC and those at lead times fall more within regions of positive GFC. For all EKE maxima, the contribution to EKE growth from GFC exceeds both the baroclinic and barotropic conversion terms. These results are consistent with those in the previous section and confirm the importance of the GFC in the decay of downstream waves and amplification of upstream waves.

![Fig. 5. As in Fig. 4, but for lag = 24 h.](image-url)
The EKE budget averaged between 15° and 20°N reveals a slightly different picture (Fig. 10). According to Fig. 4, EKE maxima strengthen through barotropic conversion east of about 20°W and weaken through barotropic conversion west of this longitude (Fig. 10b). These results are consistent with Leroux et al. (2010), who showed that the northern flank of the exit region of the AEJ is enhanced during periods of stronger AEW activity and suggests the AEJ is strengthening through barotropic conversion from the AEWs. Also consistent with previous studies, the baroclinic term (Fig. 10a) is stronger for the 15°-20°N AEWs than for the 6°-14°N AEWs (Kiladis et al. 2006). In contrast to EKE centers along 6°-14°N, EKE maxima spend most of their lives in regions of geopotential flux divergence (Fig. 10c). Once again, strong baroclinic EKE generation is often counteracted by divergent geopotential fluxes (Orlanski and Sheldon 1995). These results suggest that GFC may be less important to wave packet evolution for northern track AEWs as compared with the southern-track AEWs.

b. Vertical distribution of fluxes

The results in previous sections have shown that energy dispersion in AEWs is directed preferentially upstream (eastward) and is mediated by ageostrophic geopotential fluxes. To address this upstream energy dispersion, the vertical distribution of ageostrophic geopotential flux and its relationship to the background PV gradient is examined.

The climatological meridional PV gradient over northern Africa (contours in Fig. 11) has a complicated pattern. The negative PV gradient associated with the anticyclonic shear side of the AEJ is surrounded on all sides by a positive PV gradient. The barotropic Rossby wave dispersion relation predicts westward phase velocity and eastward group velocity within a positive meridional PV gradient and the reverse within a negative gradient. Since AEWs extend into both gradients, they could theoretically have both eastward and westward group velocity. To understand why the eastward group velocity dominates, the zonal component of the ageostrophic geopotential flux is averaged over one wavelength (or, equivalently from lag = -42 to 42 h of the regression) on a latitude–height cross section along the longitude of the base point (15°W). This average (shaded) is superimposed on the PV gradient (contoured) in Fig. 11.

Strong westward ageostrophic geopotential fluxes occur between 12° and 18°N near 650 hPa. This region coincides with the strongest negative PV gradient and the altitude at which AEWs have their largest amplitude.
Moving away from this region, the fluxes at 650 hPa become eastward near the locations at which the PV gradient goes to zero. This pattern implies that AEWs at 650 hPa have both eastward and westward group velocity depending on their latitude and is consistent with the view that barotropic instability involves the mutual interaction of two counterpropagating Rossby edge waves in the horizontal plane (e.g., Hoskins et al. 1985). However, unlike the situation at 650 hPa, the change in direction does not occur exactly where the PV gradient goes to zero. This discrepancy may result from the large changes that occur to the PV field due to the diurnal cycle of dry convection, which probably leads to large changes in the height of the PV gradient reversal.

Integrating the ageostrophic geopotential fluxes with respect to height (e.g., Fig. 4) leads to a large cancellation between the eastward and westward fluxes between 15° and 20°N. As a result, the net energy propagation is close to zero in this latitude band. To the south and north of this region, the integrated ageostrophic fluxes are eastward, with the largest values to the south of the AEJ. The net result is an eastward group velocity.

c. Potential vorticity perspective

The PV framework provides additional insight regarding the group propagation of AEWs. Because we are most interested in the phase and group velocity relative to the flow, we plot only the advection of the PV field by the perturbation wind \((-\mathbf{v}\cdot\nabla q\)) for reference. Fig. 12 shows the mean 318-K PV and PV gradient.

![Fig. 7. Residual of the EKE budget (10^{-6} W kg^{-1}, shaded) and the 950–350-hPa integrated EKE (J kg^{-1}, contours) for lag = 0 h.](image)

![Fig. 8. Regressed TRMM precipitation rate (mm h^{-1}, shaded); vertically integrated EKE (J kg^{-1}) and 650-hPa perturbation winds (m s^{-1}, vectors). All fields are based on regressions against 2–10-day bandpass-filtered meridional wind at the 10°N, 15°W base point.](image)
the strip of locally high PV along 10°N, which can be attributed to deep, moist convection, and the locally low PV in the Sahara, which can be attributed to shallow, dry convection. This PV configuration leads to a large positive meridional PV gradient south of 10°N and a negative PV gradient north of 10°N.

Figure 13 shows PV anomalies (shaded) and PV advection by the perturbation wind at 318 K (contours). These fields are shown every 24 h from lag = −24 to 48 h. The thick black contour marks the location of zero meridional PV gradient. At lag = −24 h, the PV field has a positive anomaly centered near the coast at 10°N, 10°W straddled by two negative anomalies to the east and west (Fig. 13a). As a result of the gradient reversal, the advection changes signs along 10°N on either side of the zero PV gradient contour. Thus, in northerly flow to the west of the positive PV center, PV advection is negative to the north of 10°N, while it is positive to the south of 10°N. Based on the sign of the PV advection, the relative phase velocity north of 10°N is upstream (eastward) and the relative group velocity is downstream (westward). Conversely, to the south of 10°N, the relative phase velocity is downstream and the relative group velocity upstream. Although this analysis suggests that
both upstream and downstream group velocity exist, the upstream group velocity dominates, since the magnitudes of the advection south of 10°N are much larger. This conclusion is consistent with Figs. 4–6 and suggests that the positive large PV gradient south of the ITCZ (Fig. 12) is an important region for upstream wave dispersion.

The growth or decay of individual PV anomalies through energy dispersion is manifested in the zonal asymmetry of the advection field. Nielsen-Gammon and Lefevre (1996) provide a useful schematic of this process in their Fig. 2. For example, in Fig. 13a, the negative PV anomaly near the prime meridian has stronger negative advective tendencies ahead of it and negligible tendencies behind it. These stronger negative tendencies result from the combined flow from the much stronger downstream positive anomaly and the weaker upstream negative anomaly. Without compensating positive tendencies behind it, these negative tendencies will amplify the negative PV anomaly. Based on Fig. 13, the amplitude of this anomaly increases from about −0.03 to −0.05 potential vorticity units (PVU) from lag = −24 to 0 h. This increase can be easily accounted for by the −0.03 to −0.04 PVU day⁻¹ tendency ahead of the negative anomaly over the course of a day (Figs. 13a,b). Diabatic PV tendencies should be negligible for this anomaly, since it remains located within the anomalous dry phase of the AEW (Figs. 8a,b).
By lag = 24 h, the upstream negative anomaly has reached the coast and is near its maximum amplitude, whereas the positive anomaly to its west has weakened slightly. This time evolution is consistent with an upstream group velocity relative to the flow. The advection pattern around the negative anomaly is also generating or at least reinforcing a new positive anomaly near 5°W. Though the explanation for the amplification of this new anomaly is complicated by its collocation with active convection (Fig. 8c), which would also cause positive PV tendencies, the growth rates due to advection are sufficient to sustain this anomaly and the active convection is located slightly to the north of the center of the anomaly. Thus, energy dispersion likely explains at least some of its growth.

By lag = 48 h, this new positive PV anomaly reaches its maximum amplitude near the West African coast (Fig. 13d). It appears to be growing through both energy dispersion and diabatic heating. Meanwhile, the advection pattern around the negative anomaly in the eastern Atlantic has become more symmetric compared to previous times and this suggests that its growth via energy dispersion is lessening.

The PV analysis also suggests that AEWs near and off the West African coast have a slightly southward component to their group velocity. The ageostrophic geopotential flux vectors shown in Figs. 5 and 6 also reveal a southward component of energy dispersion. The reason for this becomes apparent through Fig. 13. Strong PV anomalies near the coast of West Africa induce the strongest PV tendencies of the opposite sign to their south and east, where the positive meridional PV gradient is the strongest (Fig. 12). Thus, new PV centers of opposite sign will amplify preferentially to the south and east of existing ones.

4. Discussion and conclusions

The results presented here show that AEW wave packets are clearly associated with energy dispersion as diagnosed through group velocity. The AEJ is associated with a reversal in the meridional PV gradient and thus both upstream (eastward) and downstream (westward) group velocity are possible. However, the results show that upstream group velocity associated with the poleward PV gradient dominates wave packet evolution. This behavior is observable in Hovmöller plots and is consistent with the direction of the ageostrophic geopotential fluxes and the advective PV tendencies induced by the PV anomalies associated with AEWs. This phenomenon is similar to that of downstream development in extratropical cyclones and provides a mechanism whereby new AEWs can form upstream of existing AEWs.

An upstream group velocity may explain the tendency for AEWs to travel in groups of about three waves as seen in Fig. 1 and also noted by Pytharoulis and Thorncroft (1999). Once a strong AEW has been initiated, an eastward group velocity would favor several more AEWs developing upstream. What limits this process to about three waves is unclear, but it may be linked to the tendency of strong AEW activity to stabilize the AEJ by reducing its vertical and horizontal shear. This behavior can also be seen in the results of Thorncroft et al. (2008). In their simulation, when a train of AEWs is triggered by heating in eastern Africa, subsequent AEW troughs and ridges become stronger than the initial AEW instigated by the heating. Several of their simulations also show an eastward dispersion of energy as far east as the Indian Ocean. Thus, some of the observed short-term...
intermittency of AEWs may be mediated by their tendency to travel in distinct wave packets.

Upstream energy propagation may have implications for AEW track dynamics. In their study on the dynamics of storm tracks in the midlatitudes, Chang and Orlanski (1993) found that the downstream radiation of EKE through ageostrophic geopotential fluxes extended the storm track from regions of higher baroclinicity to regions of lower baroclinicity. By a similar argument, it could be speculated that this process limits the track of AEWs on the downstream (western) end of their storm track. This speculation is consistent with Fig. 6, which shows strong EKE decay through geopotential flux divergence within the EKE center located over the eastern Atlantic. It is also consistent with Fig. 3, which shows that the meridional wind signal following the group velocity is more coherent on the upstream side of the regression base point for AEWs in the eastern Atlantic. Indeed, it is well established that AEWs typically decay upon exiting the West African coast, where they reach their maximum intensity (e.g., Kiladis et al. 2006; Hall et al. 2006). However, other factors should not be overlooked. The thermodynamic environment over the ocean is less favorable for sustaining convection, and the PV gradient reversal associated with the AEJ weakens. To explore which of these processes dominates, a more sophisticated EKE budget or a numerical model capable of detailed process-based sensitivity studies will be required. A large loss of energy through dispersion may also be relevant to tropical cyclogenesis and to the survival of AEWs over the Atlantic in general. If high-amplitude AEWs lose a significant amount of energy to dispersion, then it may be vital for cyclogenesis that they develop a closed circulation (e.g., Dunkerton et al. 2009) to minimize energy dispersion through Rossby waves.

Based on the results presented here, and drawing on previous studies, we hypothesize the following life cycle for a period of enhanced AEW activity:

- An AEW wave train is initiated during a period when the AEJ is favorable for wave activity (e.g., Leroux and Hall 2009). The initiation may result from convection in eastern Africa (e.g., Thorncroft et al. 2008; Mekonnen et al. 2006; Leroux et al. 2010) or forcing from the North Atlantic storm track (Leroux et al. 2011). Although geopotential flux convergence cannot initiate new AEWs unless older AEWs already exist, maxima in AEW activity may still be preceded by weaker AEWs dispersing energy upstream, as suggested by Figs. 2, 3, 9c. Nevertheless, evidence from previous studies suggests that an initial convective or mid-latitude trigger is important.
- Once sufficient AEW activity is established, strong ageostrophic geopotential fluxes generate or intensify...

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**Fig. 13.** PV anomalies (PVU, shaded), advective tendencies (contours, interval of 0.01 PVU day$^{-1}$), and winds (m s$^{-1}$, vectors) at 318 K for lag = (a) −24, (b) 0, (c) 24, and (d) 48 h. Thick black line marks the zero meridional gradient of the mean 318-K PV field.
AEWs upstream while causing them to decay downstream (Fig. 9). This process can maintain AEWs through several wave periods (Fig. 1), and it may be a mechanism for sustaining AEWs during periods of increased AEW activity.

- After several wave periods, AEW activity begins to decay. We speculate that after the passage of several AEWs, the AEJ is too stable to support additional AEWs.

The results presented here indicate that group velocity is a useful conceptual framework for describing the evolution of AEWs. However, additional effort is needed to assess how well the behavior of individual groups of AEWs matches that of the statistical description based on regressed fields. It is also important to examine the impact of cumulus convection on the group and phase velocities of the waves and the structure of the AEW storm track. An additional aspect of interest is the potential for convergence of group velocity in a spatially varying background flow that can impact AEW evolution and provide a pathway to TC genesis (e.g., Webster and Chang 1988; Kuo et al. 2000; Done et al. 2010).

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