

African Easterly Wave Variability and Its Relationship to Atlantic Tropical Cyclone Activity

CHRIS THORNCROFT

Department of Meteorology, University of Reading, Reading, United Kingdom

KEVIN HODGES

Environmental Systems Science Centre, University of Reading, Reading, United Kingdom

(Manuscript received 30 November 1999, in final form 23 May 2000)

ABSTRACT

Automatic tracking of vorticity centers in European Centre for Medium-Range Weather Forecasts analyses has been used to develop a 20-yr climatology of African easterly wave activity. The tracking statistics at 600 and 850 mb confirm the complicated easterly wave structures present over the African continent. The rainy zone equatorward of 15°N is dominated by 600-mb activity, and the much drier Saharan region poleward of 15°N is more dominated by 850-mb activity. Over the Atlantic Ocean there is just one storm track with the 600- and 850-mb wave activity collocated. Based on growth/decay and genesis statistics, it appears that the 850-mb waves poleward of 15°N over land generally do not get involved with the equatorward storm track over the ocean. Instead, there appears to be significant development of 850-mb activity at the West African coast in the rainy zone around (10°N, 10°W), which, it is proposed, is associated with latent heat release.

Based on the tracking statistics, it has been shown that there is marked interannual variability in African easterly wave (AEW) activity. It is especially marked at the 850-mb level at the West African coast between about 10° and 15°N, where the coefficient of variation is 0.29. For the period between 1985 and 1998, a notable positive correlation is seen between this AEW activity and Atlantic tropical cyclone activity. This correlation is particularly strong for the postreanalysis period between 1994 and 1998. This result suggests that Atlantic tropical cyclone activity may be influenced by the number of AEWs leaving the West African coast, which have significant low-level amplitudes, and not simply by the total number of AEWs.

1. Introduction

African easterly waves (AEWs) are an important part of the West African and tropical Atlantic climate. They are known both to modulate the daily rainfall over West Africa (e.g., Reed et al. 1977) and to initiate most Atlantic tropical cyclones (e.g., Landsea et al. 1998). Despite their importance, very little is known about the detailed aspects of their life cycle, including genesis, growth, structural developments along their track, and decay.

The main aim of this paper is to present a 20-yr AEW climatology based on the period 1979–98, including a description of the seasonal cycle and interannual variability of AEW activity. The analysis is based on the European Centre for Medium-Range Weather Forecasts (ECMWF) reanalysis dataset (available from 1979 to

1993) together with ECMWF operational analyses (between 1994 and 1998). We will objectively analyze the AEWs using the automatic tracking technique developed by Hodges (1995). We have previously used this technique to track mesoscale convective systems over West Africa (Hodges and Thorncroft 1997).

Previous analysis of AEWs using operational analyses has mainly used statistical techniques to diagnose the AEW activity (e.g., Reed et al. 1988a; Lau and Lau 1990; Fyfe 1999; Ceron and Gueremy 1999). These papers have diagnosed different aspects of the AEW activity, including the different regions of AEW activity north and south of the African easterly jet. In contrast, a different perspective of the nature of AEW activity was obtained by Reed et al. (1988b), who manually tracked AEWs between August and September in 1985. This analysis was able to show the tracks of individual systems moving over Africa, over the Atlantic Ocean, and, in some cases, recurving as tropical cyclones. The analysis presented in this paper expands on the analysis of Reed et al. (1988b). By making use of the automatic tracking method, it is now possible to consider the AEW tracks in more years and to generate important AEW

Corresponding author address: Dr. Chris Thorncroft, Dept. of Meteorology, University of Reading, Earley Gate, P.O. Box 243, Reading RG6 6BB, United Kingdom.
E-mail: c.d.thorncroft@reading.ac.uk

statistics related to their genesis, growth, and decay. It should be recognized that results using the tracking technique offer a different but complimentary view to results using statistical techniques. A statistical method applied to the geopotential field, for example, will give a smoother view of AEW activity than a tracking method used to follow vorticity centers within the AEWs and it is important to recognize this. The present study focuses on the positive relative vorticity centers that exist north and south of the easterly jet both because of the new insight gained over Africa and their relevance as potential precursors for tropical cyclones. It should be noted that while the analyses considered here are derived from following just positive relative vorticity centers, the resulting tracks will be referred to as AEW tracks.

An important motivation for the study of AEW variability relates to the marked interannual variability of West African rainfall (e.g., Rowell et al. 1995). Since AEWs are the dominant synoptic systems in the region, it is important to investigate whether the AEWs have an important role to play in determining the interannual rainfall variability. Indeed, it is still unclear whether AEWs have an active or passive role in this.

A further motivation for considering the interannual variability of AEWs concerns the relationship between AEWs and Atlantic tropical cyclones. Several authors have suggested that the positive correlation between Atlantic tropical cyclone activity and West Sahel rainfall might be due to the AEW variability (e.g., Reed 1988; Gray 1990; Landsea and Gray 1992). Two aspects of AEW variability that have been raised in this regard are their frequency and their intensity. Regarding frequency, it seems entirely plausible that if there were more potential precursors this would increase the probability of tropical cyclogenesis given the same favorable large-scale conditions such as warm SSTs. However, Avila and Pasch (1992) suggested that AEW variability is weak, with 59 waves a year and a standard deviation of only 4.4. It should be remembered though that this study was mainly based on satellite data and therefore diagnosing whether a convective system is an AEW or not is quite subjective. We suggest that a dynamical measure may help to diagnose the AEW activity more objectively and may help us in our understanding of AEW and tropical cyclone variability. In contrast to Avila and Pasch (1992), for example, Thorncroft and Rowell (1998) found marked interannual variability in AEW activity in a long GCM integration using filtered meridional wind variance as a diagnostic.

Reed (1988) was probably the first to suggest that, rather than variability in frequency, variability in AEW intensity could be more important in determining the tropical cyclone variability. This is also a plausible hypothesis and is consistent with the known requirements for a finite-amplitude precursor for tropical cyclogenesis (e.g., Gray 1979; Emanuel 1989) and/or the conducive effect of a high vorticity background for genesis (e.g., Gray 1979; Schubert et al. 1980). However, care needs

to be taken when defining what is meant by intensity of an AEW because of its complicated structure. Recently, Pytharoulis and Thorncroft (1999, hereinafter PT) have confirmed that AEWs may be multicentered, with maxima in vorticity both near the African easterly jet level around 600–700 mb in the rainy zone and also around 850 mb and lower, north of the rainy zone. In any examination of AEW variability the complicated nature of AEW structures must be considered. It follows therefore that as well as variability in number and intensity, variability in structure may also be important in determining tropical cyclone variability.

This paper presents the AEW variability together with an assessment of the relationship between the waves and the tropical cyclones. A later paper will examine more closely the relationship between the wave variability and the West African monsoon, including the rainfall and African easterly jet (AEJ) variability.

The paper is organized as follows. Section 2 will outline the tracking method used. Section 3 will show examples of the tracks obtained from this technique by comparing the 1995 and 1994 seasons. In section 4 the AEW climatology is presented based on the tracking statistics for 1979–98, including a description of the seasonal cycle. In section 5 we focus on the interannual variability of AEW activity and discuss whether this is related to tropical cyclone activity. The results are summarized and discussed further in section 6.

2. Tracking technique

Reed et al. (1988b) highlighted the usefulness of using vorticity to manually identify and track AEWs in ECMWF operational analyses. Here we apply automated methods of identification and tracking of AEWs to such data. The data used for this study are the ECMWF reanalysis for 1979–93 (Gibson et al. 1997) extended with operational analyses to 1998. We consider only the May–October period, to focus on the West African monsoon rainy season, and use 6-hourly initialized analyses. Results are only presented for the 600- and 850-mb levels, because these are able to clearly distinguish the AEW activity at the jet level in the rainy zone from that at lower levels poleward of the main rainy zone. A spectral resolution of T42 was used, which is adequate to resolve the synoptic-scale AEWs, the main concern here. Also, no advantage was found to using higher-resolution analyses. Generally, the higher-resolution analyses appeared to be “noisier” versions of the T42 analyses, which may be consistent with lack of adequate high-resolution observations to assimilate.

The method used to identify and track the waves is that of Hodges (1995). This method identifies the waves as maxima in the thresholded relative vorticity field (a threshold of $+0.5 \times 10^{-5} \text{ s}^{-1}$ is used) and tracks them using a method that minimizes a cost function for the motion coherence subject to constraints on the motion. The cost function is defined in terms of local changes

in direction and speed. This results in the minimal set of smoothest tracks. The constraints are for the maximum displacement distance in a time step and the local track smoothness and are applied adaptively (Hodges 1999). It should be noted that we are only detecting systems that have closed vorticity contours with values of at least $+0.5 \times 10^{-5} \text{ s}^{-1}$ with this approach. In a sense the tracking, as it is applied here, is acting as a filter on the waves by excluding the weaker waves that do not support such coherent high vorticity centers.

The track ensembles are also filtered to remove all systems with lifetimes of less than 2 days and that travel less than 10° ($\sim 1000 \text{ km}$) so that only the coherent systems are retained for statistical analysis. Statistics are computed from the combined 20-yr track ensemble using the kernel methods described by Hodges (1996) for track and genesis densities and mean growth/decay rates. The densities are scaled to number densities per unit area from probability density distributions, where the unit area is equivalent to a 5° spherical cap ($\sim 10^6 \text{ km}^2$).¹ It should be noted that the genesis density may include situations where systems fall below the intensity threshold and then reintensify. For example, the AEWs can weaken enough to make them difficult to detect in the midocean but may then reappear downstream. Also, the mean growth/decay rate statistic should be interpreted in conjunction with the track density, because the values will be more reliable where the density is higher.

An example of a single track is shown in Fig. 1a of Hurricane Luis from the 1995 hurricane season. The track of Luis from when it first became a tropical storm, analyzed by the National Hurricane Center (NHC, now known as the Tropical Prediction Center) is included (gray circles) alongside the track identified using 850-mb vorticity from the ECMWF data (black circles). Overall the tropical cyclone track is well represented in the ECMWF data. However, it is important to notice that the tracking system identifies the AEW associated with the storm farther east than is indicated by NHC, consistent with the tropical cyclone developing from a weaker AEW system. Also in Fig. 1b are some typical tracks from 1995 based on the 600-mb vorticity centers, which were systems that did not become tropical cyclones. Note that while these systems were tracked from the African continent into the mid-Atlantic they could not be tracked into the Caribbean, suggesting that these systems weaken as they move westward.

3. African easterly wave tracks for 1994 and 1995

It has recently been confirmed by PT that AEWs can have quite complicated structures over Africa, with

¹ Note that in Hodges and Thorncroft (1997) the track density presented was calculated differently and was not a probability density distribution. Here the track density is calculated as a pdf and then scaled as number density and so integrates to the total number of systems.

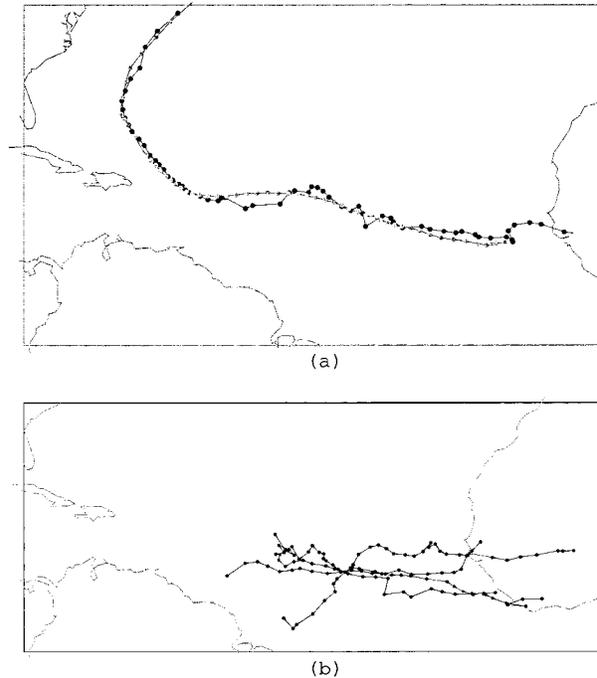


FIG. 1. (a) Hurricane Luis (1995) from the tracking analysis performed on the ECMWF analyses of 850-mb relative vorticity and the best track from the National Hurricane Center (NHC). The dots indicate the position every 6 h: the dark dots are for ECMWF data and the light dots are for the NHC best track; (b) a selection of typical tracks at 600 mb showing systems that did not develop into tropical cyclones.

maximum amplitudes typically seen in two regions: at around 600 mb equatorward of about 15°N and also at low levels poleward of 15°N . This is consistent with AEWs growing through a mixed barotropic–baroclinic instability mechanism in association with negative meridional gradients in potential vorticity at the AEJ level and positive meridional gradients in temperature at the surface (e.g., Burpee 1972). Recognizing this, we therefore present AEW tracks at 600 and 850 mb in order to diagnose the AEW structures in both regions.

Figures 2 and 3 show the May–October analyzed tracks for 1995 and 1994, respectively. These are presented in order to further illustrate the output from the automatic tracking analysis and also to compare the nature of the AEWs in two contrasting Atlantic tropical cyclone seasons. The 1995 season was an extremely active year with 19 named storms (Landsea et al. 1998). The 1994 season, in contrast, was a very inactive year with only seven named storms.

In 1995 the AEW tracks can clearly be seen at 600 mb over west Africa equatorward of about 15°N . These tracks lead into a more active storm track region over the Atlantic, with many tracks subsequently recurving poleward before reaching the Caribbean. At 850 mb the tracks over the ocean are similar to those at 600 mb, except that there are generally fewer of them, and the recurving tropical cyclones are more clearly distin-

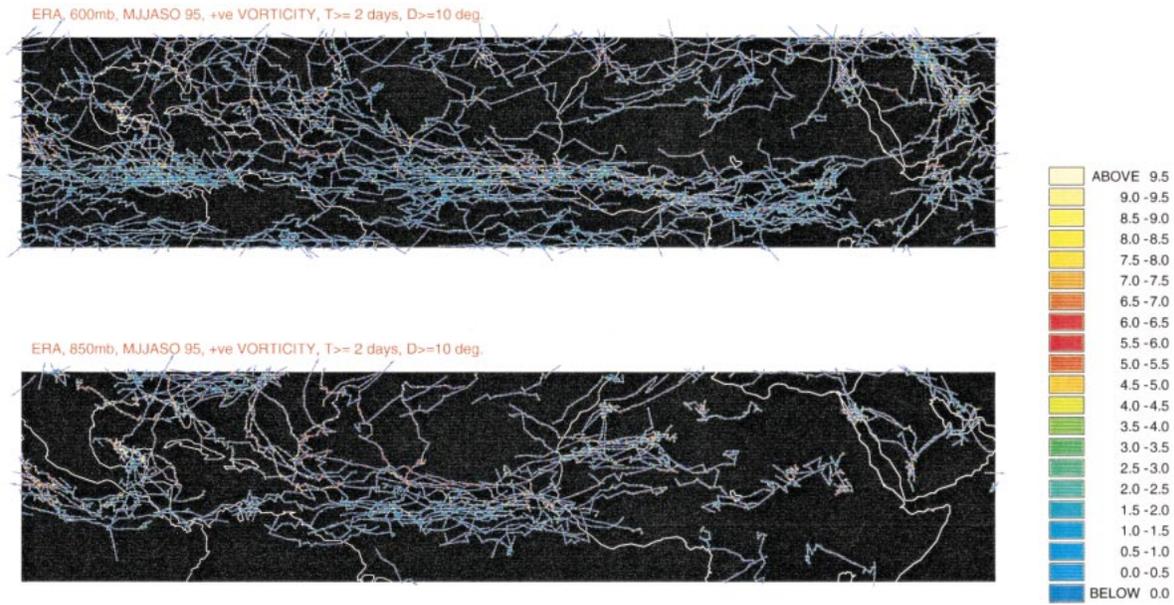


Fig. 2. Positive relative vorticity tracks for May–Oct 1995 at 600 and 850 mb.

guished consistent with their large amplitudes and warm core structures. Larger differences occur over the land though, where the main AEW storm track is poleward of 15°N. Indeed, at 850 mb very few tracks exist equatorward of about 15°N. This is consistent with previously analyzed AEWs in this region (e.g., Reed et al. 1977), which tend to have amplitudes maximizing near the level of the African easterly jet, around 600 mb with weaker amplitudes below. It also further confirms the dominance of low-level structures poleward of the AEJ

in the region of the low-level temperature gradient on the fringes of the Sahara (cf. PT).

In 1994 the AEW tracks at 600 mb are very similar to those in 1995 over Africa and the Atlantic, although consistent with the weaker tropical cyclone activity that year, there are fewer recurring tracks near the Caribbean. Bigger differences between the two years are evident at 850 mb. Consistent with the weaker tropical cyclone activity, there are fewer and weaker tracks in the tropical Atlantic. It is particularly interesting, how-

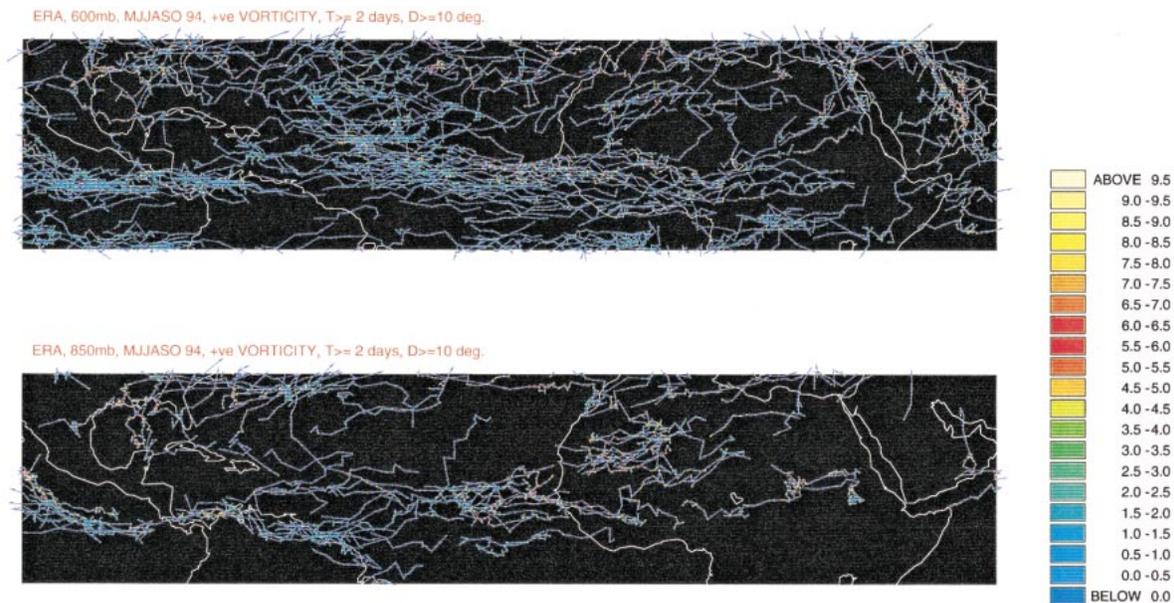


Fig. 3. Positive relative vorticity tracks for May–Oct 1994 at 600 and 850 mb.

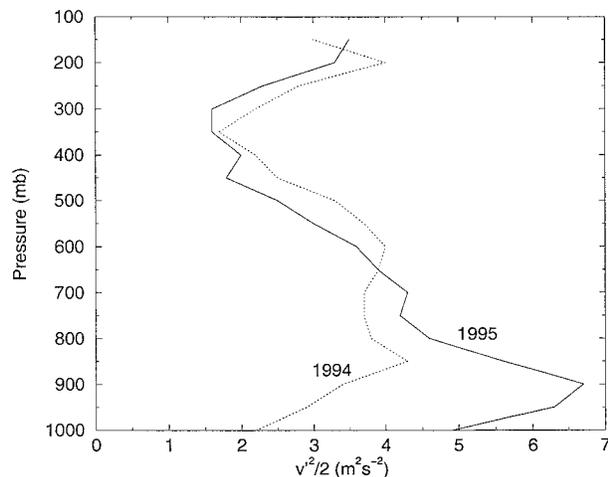


FIG. 4. Vertical profile of the Jun–Sep mean $v'^2/2$ for Dakar (15°N , 17°W), where v' is the 2.5–6-day bandpass-filtered meridional wind. The solid line is for 1995 and the dashed line is for 1994. The figure was provided by Pytharoulis and is based on analysis presented in Pytharoulis (1999).

ever, that this weaker 850-mb activity is also evident at the West African coastline. This suggests that perhaps one important difference between the two years is in the different levels of AEW activity coming out of Africa, particularly at low levels. Analysis of filtered radiosonde winds at Dakar (15°N , 17°W) in 1994 and 1995 (Fig. 4) is perhaps consistent with this, although Dakar is a little poleward of where most of the 850-mb tracks leave the West African coast. The idea that low-level AEW amplitudes could be important for influencing tropical cyclone activity is consistent with the known requirement for a finite-amplitude low-level precursor for tropical cyclogenesis (e.g., Gray 1979; Emanuel 1989). This hypothesis will be returned to in the next section where we present AEW tracking statistics based on all the years in this study.

4. African easterly wave tracking statistics climatology (1979–98)

a. Motivation

Here we present a description and interpretation of the 20-yr climatological AEW tracking statistics. From this we can have a better perspective of the typical AEW life cycle, including initiation, growth and decay, and basic storm track information. As well as providing this climatological view, it will give us the necessary background with which to consider the interannual variability of AEW activity in section 5.

b. 600 mb

1) TRACK DENSITY

The 600-mb track density (Fig. 5) shows a very clearly defined storm track around 10° – 15°N , starting over

east Africa and stretching out westward over the Atlantic. There is also a hint that it joins with the storm track seen in the east Pacific. Closer inspection of individual tracks also confirms that some tracks that start in the Caribbean track across the land and into the Pacific in agreement with Molinari et al. (1997). Just east of the Caribbean there is some evidence for recurving storms, but since recurving tropical cyclones only make up a small percentage of all systems tracked and do not always recurve at the same longitude, this feature is weaker. Notable peaks appear just downstream of the West African coastline around 20°W and also downstream of Central America in the east Pacific. Weaker activity exists over the African continent. This is consistent with weaker waves early in their life cycle and previous analysis of radiosonde data by Albignat and Reed (1980), who found very weak AEW activity east of 10°E . It should also be remembered though, that the tracking scheme tracks only systems with closed vorticity contours greater than or equal to $0.5 \times 10^{-5} \text{ s}^{-1}$. This may mean that weak AEWs early in their life cycle are not tracked using this scheme and may contribute to the weaker analyzed activity over land.

The peak in AEW activity just downstream of the African continent is consistent with previous studies based on observed filtered wind variances (e.g., Albignat and Reed 1980). What the present analysis shows more clearly though is the continuation of the AEW storm track into the central Atlantic and into the tropical cyclone “main development region” (cf. Goldenberg and Shapiro 1996).

2) GENESIS DENSITY

The question of AEW genesis has been raised by several authors previously. Whereas consensus generally exists regarding the growth of AEWs through a mixed barotropic–baroclinic instability mechanism, consensus over where the AEWs are initiated has not been reached. Carlson (1969) suggested that AEWs could be initiated as far east as the Ethiopian highlands ($\sim 40^\circ\text{E}$). However, Burpee (1972) could only find AEW activity in radiosonde data as far east as 15°E . A detailed analysis on the initiation of AEWs during Phase III of the GARP Atlantic Tropical Experiment was presented by Albignat and Reed (1980). They suggested that AEWs may originate from as far east as the Red Sea (around 40°E). As noted by Albignat and Reed (1980), one problem with comparing different studies is that they have considered different periods, and so disagreement may just mean that AEW activity fluctuates in these regions.

Another reason, which will not be easy to overcome in the short term, is the sparsity of data in the likely genesis regions. Here, we can overcome the data uniformity problem but may still suffer from a problem arising from data sparsity. For this reason, we prefer to

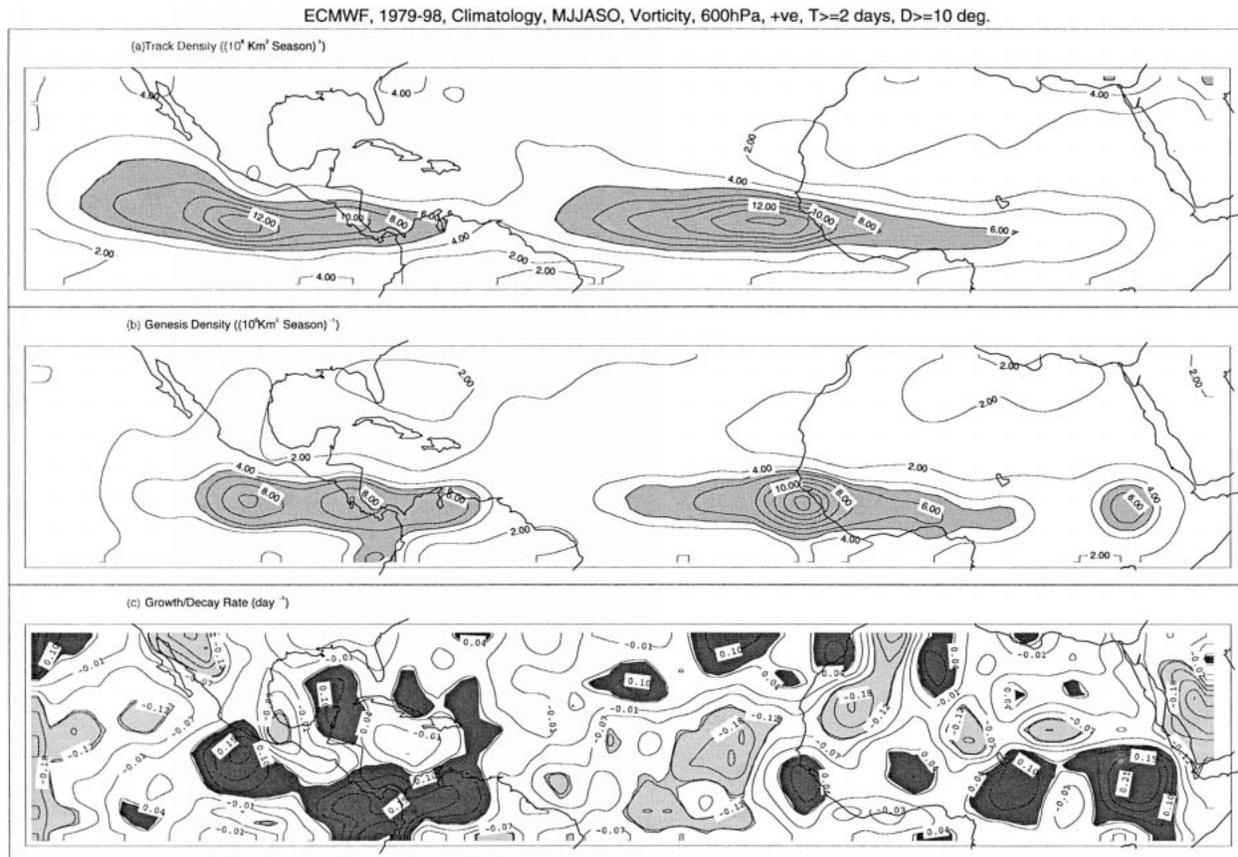


FIG. 5. Climatological tracking statistics at 600 mb based on the ERA data (1979–93) and the ECMWF analyses (1994–98). (a) Track density scaled to number density per unit area ($\sim 10^6 \text{ km}^2$) per season (MJJASO), shading for values greater than 6. (b) Genesis density per unit area ($\sim 10^6 \text{ km}^2$) per season (MJJASO), shading for values greater than 5. (c) Growth and decay rates in units of per day, shading for values greater than 0.05 and less than -0.1 .

consider the average statistics of the AEW activity rather than a detailed analysis of individual AEWs.

The 600-mb genesis density based on the 1979–98 period in Fig. 5b shows those regions where the tracked systems are first identified. The genesis regions line up with the storm track density as expected, but there are several prominent peaks. Starting in the east, there is a peak located around (10°N , 35°E) on the western side of the Ethiopian highlands suggesting that at least some AEWs are initiated there. Several tracks can be seen to start from this region in Figs. 2a and 3a. West of this region the genesis density weakens and then starts to increase again west of about 20°E . Albigat and Reed (1980) had a similar result and suggested that this was consistent with the increased strength of the AEJ. This was also shown by PT to be a region where the reverse potential vorticity gradients begin, consistent with an increased dynamical instability.

The most pronounced genesis region is located at the West African coast, with the actual peak just over the sea. This feature is at odds with the conceptual model of an AEW, which is initiated over the continent and grows as it moves along the AEJ. The peak in genesis

at the coast could be linked to the fact that the tracking criteria are acting to filter out the weaker and more disorganized systems. The fact that there is a genesis peak at the coast could then be due to the fact that many AEWs do not reach a large enough amplitude to be tracked until they reach the coast. Alternatively, the coastal region may be a region where the AEWs experience extra development, in association with latent heat release in deep moist convection, for example. This point will be returned to below in the section on growth and decay.

3) GROWTH AND DECAY RATES

The growth and decay rates at 600 mb are shown in Fig. 5c. The striking feature here is the predominant growth over the land and decay over the ocean. The growth over the land is consistent with AEWs growing on the AEJ. The decay just downstream of the West African coast was also noted by Carlson (1969), who suggested that the decay was related to convection in the wave becoming more disorganized, perhaps in association with relatively cool SSTs there. An alternative,

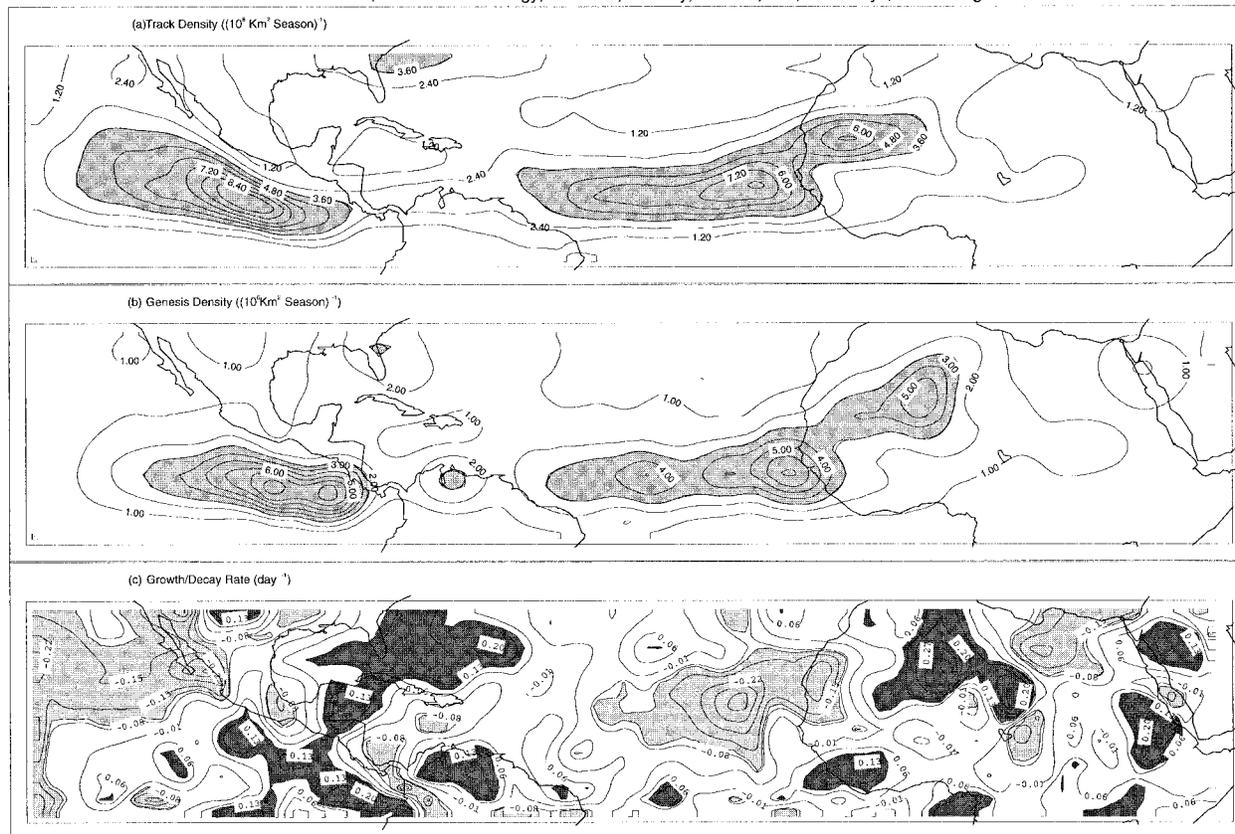
ECMWF, 1979-98 Climatology, MJJASO, Vorticity, 850hPa, +ve, $T \geq 2$ days, $D \geq 10$ deg.

FIG. 6. Climatological tracking statistics at 850 mb based on the ERA data (1979–93) and the ECMWF analyses (1994–98). (a) Track density scaled to number density per unit area ($\sim 10^6 \text{ km}^2$) per season (MJJASO), shading for values greater than 3.6. (b) Genesis density per unit area ($\sim 10^6 \text{ km}^2$) per season (MJJASO), shading for values greater than 5. (c) Growth and decay rates in units of per day, shading for values greater than 0.1 and less than -0.1 .

dynamical explanation is that whereas the AEWs over the land can make use of the unstable AEJ and the potential vorticity sign reversals that exist there, over the ocean where the sign reversal weakens (cf. PT), the AEWs may weaken through Rossby wave dispersion.

Another growth region is located over northern Venezuela and Colombia, which also coincides with a genesis peak (cf. Fig. 5b). This is a region identified by Molinari et al. (1997) as a potential vorticity sign reversal region. They suggested that AEWs could be reinvigorated there on their way into the eastern Pacific. The growth pattern shown here is consistent with this. Indeed, the growth region even continues across Central America and into the Pacific. There is also a notable peak in growth just downstream of Mexico. Mozer and Zehnder (1996a) have suggested that growth can occur in this region in association with a blocking flow around the Sierra Madre and the development of a low-level barotropically unstable region; but it is also possible that with a suitable upstream precursor such as an AEW, the warm SSTs there are able to encourage tropical cyclogenesis.

c. 850 mb

1) TRACK DENSITY

The track density at 850 mb (Fig. 6a) is similar to that at 600 mb over the ocean, though weaker, but there are major differences over the African continent. Consistent with PT the dominant track is poleward of 15°N and starts farther west than the 600-mb storm track. As shown by PT, these low-level waves often show coherency with the 600-mb waves equatorward of them, indicating a complicated multicentered AEW structure.

It can also be seen that the 850-mb storm track over the land is about 10° poleward of the 850-mb ocean storm track. This raises the question as to whether the waves analyzed at 850 mb over the land in the poleward storm track are the same as those seen in the oceanic storm track downstream or, since there is only very weak 850-mb AEW activity over the land, the 850-mb activity in the ocean storm track develops at the coast. From Figs. 2 and 3 it would seem that at least some of the waves track equatorward but not all.

The combined view of AEW activity depicted by

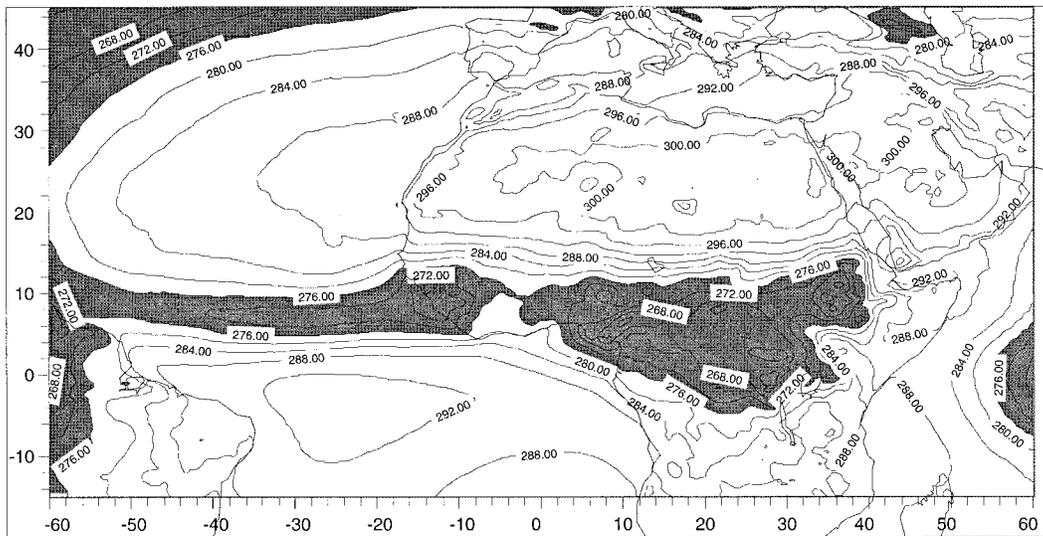


FIG. 7. Window brightness temperature from Meteosat averaged over Jun–Oct from 1984 to 1995.

Figs. 5a and 6a gives us an indication of how AEW structures change along their track. Consistent with the conceptual model proposed by PT, amplitudes at the AEJ level at around 10° – 15° N start farther east than the AEW amplitudes at low levels around 25° N on the poleward side of the AEJ. Over the ocean though, the 600- and 850-mb storm tracks become more collocated at around 10° – 15° N.

Similar to the 600-mb analysis, marked activity is found in the East Pacific consistent with the known cyclone activity there.

2) GENESIS DENSITY

Two notable peaks in genesis density exist, one poleward of 15° N over the land and one equatorward of this, just downstream of the West African coast (Fig. 6b). The first peak occurs in the region of the 850-mb storm track over the land discussed above. This is consistent with AEWs developing on the low-level temperature gradient, which maximizes in this region, but since this is also just downstream of the Hoggar mountains (25° N, 10° E), the possibly important role of orographic processes should also be considered (cf. Mozer and Zehnder 1996b). The genesis peak just off the coast is suggestive of the fact that the 850-mb AEWs generally develop there rather than track southwestward from the poleward track. There is further evidence for this in the growth/decay statistics shown below.

3) GROWTH AND DECAY RATES

The growth and decay rates shown in Fig. 6c shed some important light on the typical AEW life cycle. The low-level waves that are initiated around (25° N, 10° E) move westward and grow (indicated by the dark shading). Just downstream of this, and before they reach the

coast they generally start to decay (indicated by the light shading). This growth–decay dipole present around 20° N suggests that the low-level waves generated on the poleward side of the AEJ generally decay and do not get involved with the AEW structures equatorward of the AEJ. This does not mean that low-level waves never track equatorward. For example, a few tracks can be seen moving southwestward after leaving the coast in 1995 (Fig. 2b).

This result has important consequences for the way we interpret the 850-mb AEWs tracked over the ocean between 10° and 15° N. If most of these AEWs do not originate from the storm track poleward of the AEJ they must therefore be generated near the coast. This is consistent with the genesis maximum seen in Fig. 6c but also with the marked growth region (indicated by dark shading), which peaks at the coast equatorward of 15° N. This is a marked growth region, which starts just west of the Greenwich meridian. We conclude therefore that most of the 850-mb AEWs that we see over the ocean generally do not originate from the poleward storm track but instead are generated in association with the 600-mb AEWs over the land equatorward of the AEJ.

Two possibilities for this generation would be (i) as a natural consequence of the nonlinear life cycle and downward Rossby wave propagation consistent with Thorncroft and Hoskins (1994) or (ii) in association with diabatic processes (cf. Kwon and Mak 1990; Thorncroft and Rowell 1998). One must also consider the possibly important role played by orography in this region. This orography may also have a strong influence in determining the climatological rainfall maximum in this region. Indeed, it should be noted that the growth region at the coast is collocated with a climatological maximum in rainfall. This is illustrated here in Fig. 7, which shows the mean June–October brightness temperature for the

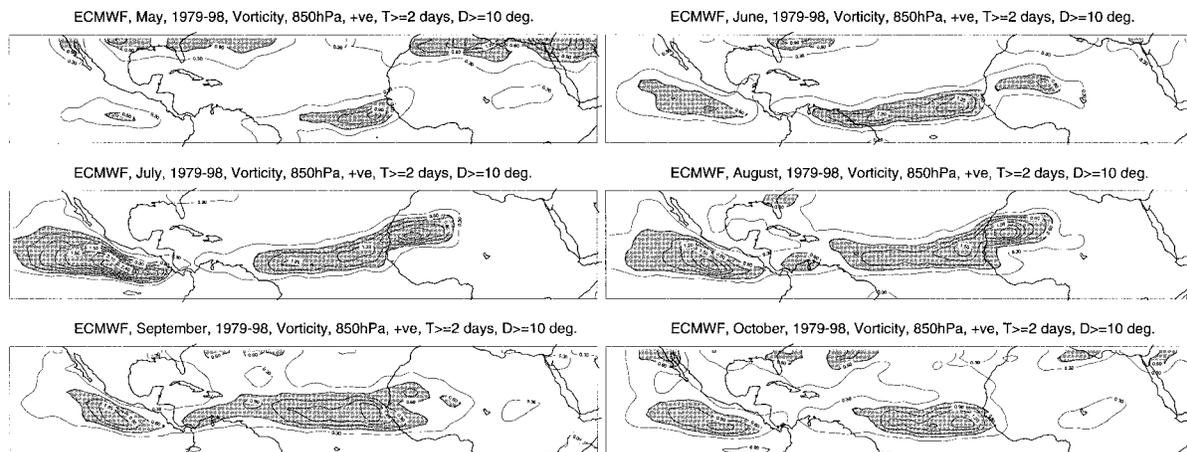


FIG. 8. Seasonal cycle track density statistic at 850 mb based on the ERA data (1979–93) and the ECMWF analyses (1994–98). Track density has been scaled to number density per unit area ($\sim 10^6 \text{ km}^2$) per month.

region based on the years 1984–95. This peak in rainfall has not been examined much previously but is likely to be associated with the elevated terrain in the region, with much of it above 1000 m, and enhanced by the land–sea contrasts. It seems very likely therefore that AEW developments at low levels are influenced by the deep moist convection that preferentially occurs in this region and that the 850-mb AEWs at the coast are diabatically generated. It should be noted that this region is equatorward of the West Sahel region, which has previously been used to relate West African rainfall variability to Atlantic tropical cyclone variability (e.g., Landsea et al. 1998).

The growth/decay rates over the ocean at 850 mb show some differences with those at 600 mb. Whereas at 600 mb there is a broad decay region, at 850 mb there is evidence for weak growth around 7°N stretching from the coast to the Caribbean. This also coincides with the position of the mean ITCZ, indicated in Fig. 7, suggesting a role for diabatic generation and/or enhancement (cf. Schubert et al. 1991).

d. Seasonal cycle of easterly wave activity

In section 5 we will present the interannual variability of AEW activity and its relationship with Atlantic tropical cyclone activity. Before this, for completeness, we will consider the seasonal cycle of AEW activity. For brevity we show only the track densities. Since we hypothesize that the AEWs at 850 mb leaving the West African coast may have an important role in tropical cyclogenesis, we concentrate mostly on this level.

1) 850 MB

Figure 8 shows the monthly mean track densities between May and October. In May at 850 mb, there is just a weak storm track indicated off the coast of West Africa and very little activity over the continent. The equa-

ward portion of the midlatitude storm track can be seen in northern Africa at this time. There is a marked increase in AEW activity in June. The poleward storm track over the continent around 20°N becomes evident for the first time and the Atlantic storm track lengthens. The increased activity over the land in June is perhaps consistent with the increased solar heating at the surface at this time and the development of a deep well-mixed boundary layer (cf. Thorncroft and Blackburn 1999), which Thorncroft (1995) argued encourages the low-level baroclinic developments on the poleward side of the AEJ. Note that by June the midlatitude storm track only weakly affects north Africa.

Between June and August the storm tracks over the continent and ocean intensify. Another notable development in August is the increased AEW activity equatorward of 15°N over the land. This equatorward track is even more pronounced in September. Also, in September, the poleward storm track over the continent weakens markedly and by October has disappeared completely. In October the only storm track evident is that over the ocean.

The increased 850-mb activity seen in August and particularly in September is perhaps consistent with the results presented by Miller and Lindzen (1992). They suggested that the increased activity in these months was due to the closer proximity between the AEJ and the moist boundary layer.

Also indicated in Fig. 8 is evidence of the recurving tropical cyclones in the western Atlantic, seen most strongly in August and September, the climatological peak for tropical cyclone activity (Landsea et al. 1998). Interestingly, there is a hint in these tracks that suggests that the recurving preferentially occurs farther west in August.

2) 600 MB

For completeness, the seasonal cycle of 600-mb track density is shown in Fig. 9. A marked seasonal

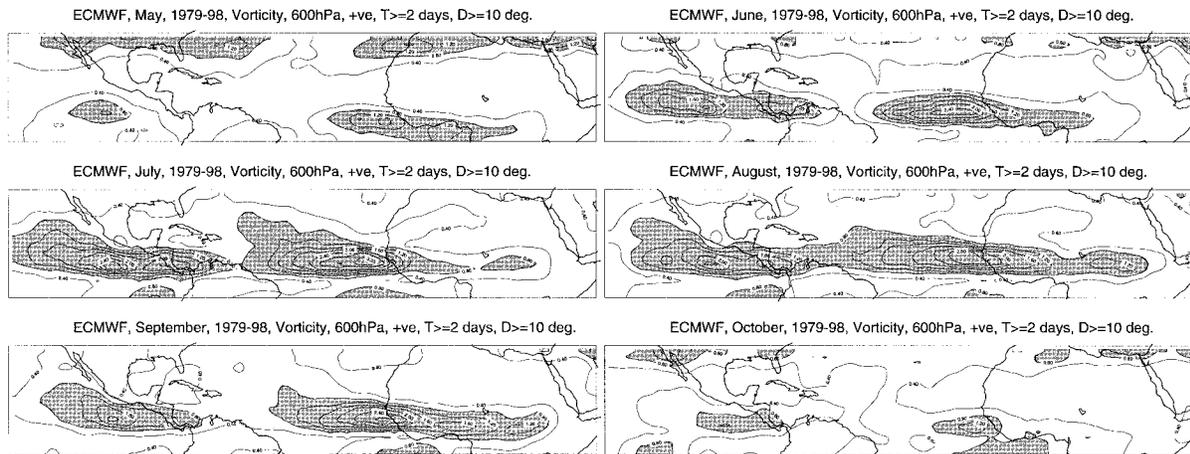


FIG. 9. Seasonal cycle track density statistic at 600 mb based on the ERA data (1979–93) and the ECMWF analyses (1994–98). Track density has been scaled to number density per unit area ($\sim 10^6 \text{ km}^2$) per month.

cycle is evident, with activity in the east Atlantic and West African regions increasing and moving poleward between May and August, followed by a weakening and rapid equatorward retreat between September and October. Over Africa the storm track is most coherent and reaches as far east as Ethiopia between July and September. One should also note the weaker track poleward of 15°N seen in July and August marking the upper portion of the structures, which peak at lower levels seen in Fig. 8.

3) TIME SERIES

In section 6, we will examine the interannual variability of AEW activity by focusing on the region between 5° and 15°N in the rainy zone. We do not consider the AEWs on the poleward side of the AEJ. This is both for brevity and because the analysis presented here sug-

gests that the poleward AEWs play only a minor role in the tropical Atlantic. We have considered three overlapping boxes (Fig. 10). Boxes 1–3 examine the AEWs over the land before reaching the coast (box 1), at the coast (box 2), and finally over the ocean after passing the coast (box 3).

For each box we simply sum the number of AEWs passing through the box. Because of the tracking criteria (see section 2, above), this may not represent all AEWs, but does represent the strongest and most coherent ones. The boxes were chosen to overlap with the region of greatest AEW activity based on the track density.

Figure 11 shows the mean seasonal cycle of tracked AEWs for each of the boxes at 600- and 850 mb. Consistent with Figs. 5a and 6a, AEW activity increases from east to west. The very marked seasonal cycle of AEW activity at both levels and for each box is clear. At 600 mb AEW activity increases from June to August, weakens slightly in September, and then decreases rapidly between September and October. The peak value of around 6 in August would suggest a period for AEWs of about 5 days. Bearing in mind that the tracking technique effectively filters out the weaker less well-defined systems this period is comparable with that typically quoted for AEWs of about 4 days.

Consistent with Figs. 8a and 9a, the mean number of AEWs tracked in each month is less at 850 than at 600 mb. This is consistent with the known cold-core structures of AEWs with weaker vorticity centers at 850 mb than at 600 mb, and the 850-mb centers falling below the threshold value used for tracking more often than at 600 mb. It should also be noted that the seasonal cycle at 850 mb differs from that at 600 mb with the peak in activity occurring 1 month later in September. Curiously, this peak coincides with the normal climatological peak in Atlantic tropical cyclone activity (Landsea et al. 1998), suggesting a possible link between 850-mb AEW activity and tropical cyclone ac-

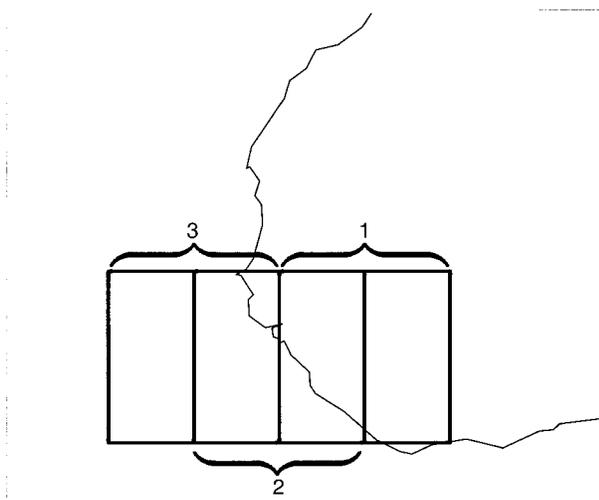


FIG. 10. The location of three boxes used for assessing the seasonal cycle and interannual variability of AEW activity.

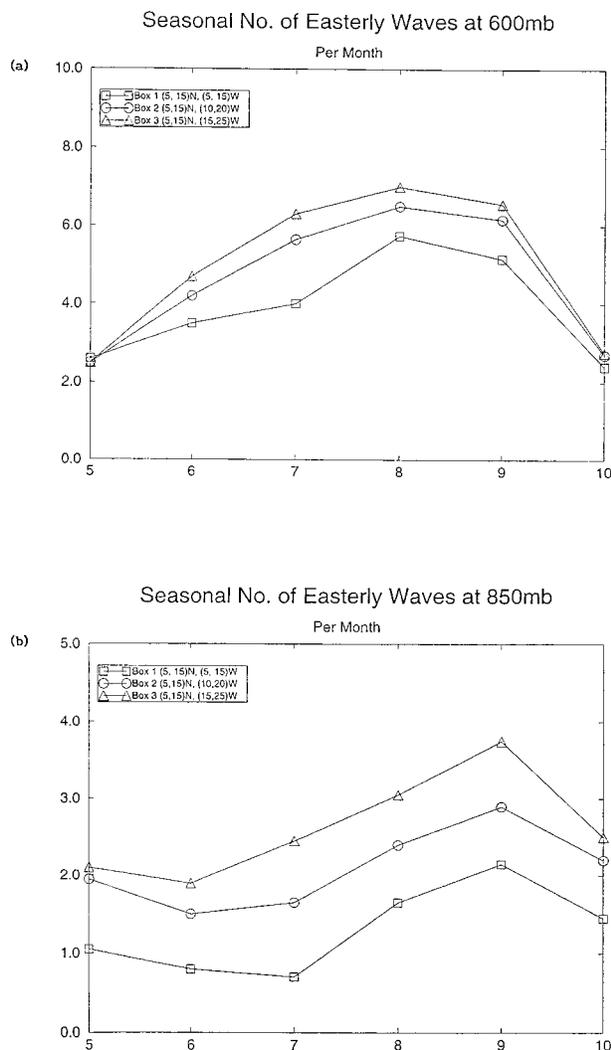


FIG. 11. Climatological seasonal cycle of the number of AEWs in the boxes shown in Fig. 10 at (a) 600 and (b) 850 mb.

tivity. This link will be investigated for interannual timescales in section 5.

5. Interannual variability

a. AEW variability

Table 1 summarizes the mean activity in each of the boxes shown in Fig. 10 together with the coefficient of variation (standard deviation divided by the mean). Interestingly, although the mean activity is lower at 850 than at 600 mb, the interannual variability at 850 mb is approximately double. The coefficient of variation varies between 0.29 and 0.41 for the 850-mb AEW activity. This compares poorly with the value of 0.07 from Avila and Pasch (1992). However, as discussed in the introduction, this earlier study is based mainly on satellite data, whereas our study is based on a dynamical measure, and so different results might have been ex-

TABLE 1. The mean and coefficient of variation (standard deviation/mean) of AEW activity at 600 and 850 mb in the three boxes in Fig. 10.

Box	Lat-long	Mean/coef. of var. 600 mb	Mean/coef. of var. 850 mb
1	(5°–15°N, 5°–15°W)	22.9/0.22	7.7/0.41
2	(5°–15°N, 10°–20°W)	27.1/0.16	12.3/0.29
3	(5°–15°N, 15°–25°W)	28.9/0.15	15.2/0.34

pected. Thorncroft and Rowell (1998) in their analysis of AEW variability in a GCM diagnosed a coefficient of variation based on 850-mb filtered meridional wind variance. In that study the coefficient varied between 0.42 and 0.50 in the west African region, which is a little larger but comparable to that seen here.

The interannual variability of AEW activity at 600 and 850 mb is plotted for box 2 in Fig. 12. The marked interannual variability in AEW activity can be seen very clearly. At 850 mb for example the peak number of AEWs tracked is 18 and the minimum is 6. At 600 mb the range is from 37 tracked in 1996 to 20 in 1985 and 1991. Note also that, consistent with the tracks presented in section 3 above, 1995 is diagnosed as a more active year than 1994 at 850 mb with 17 and 11 AEWs, respectively.

We have shown here that, based on ECMWF analyses, there is considerable interannual variability in AEW activity. We have also shown that this variability is manifested most strongly at 850 mb in the rainy zone. We hypothesize that, because tropical cyclones rely on a finite-amplitude precursor for genesis, the variability in AEW activity diagnosed here, in particular at low levels in the rainy zone, may have an impact on the variability in Atlantic tropical cyclone activity. This will be examined below.

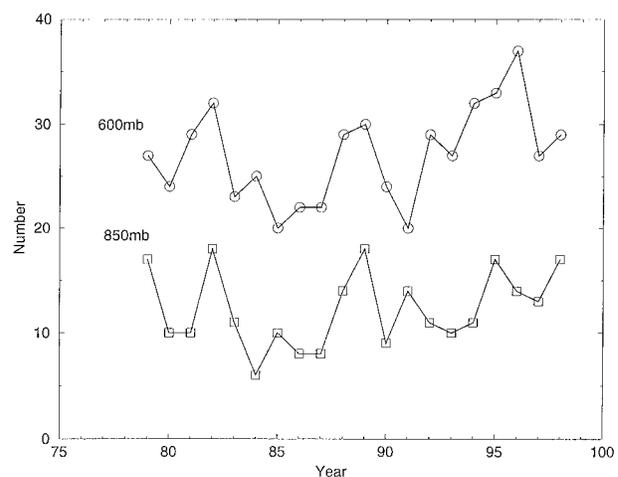


FIG. 12. Time series showing the interannual variability of the number of AEWs in box 2 (shown in Fig. 10) at 600 (circles) and 850 mb (squares); totals are based on the May–Oct period.

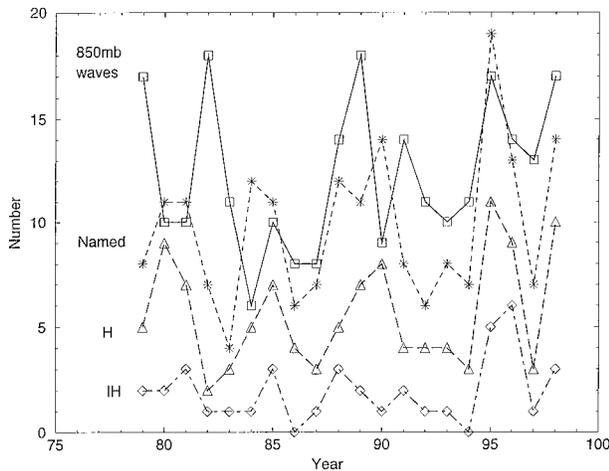


FIG. 13. Time series showing the interannual variability of the number of AEWs in box 2 (shown in Fig. 10) at 850 mb (squares) based on the May–Oct period, together with the number of named storms (stars), hurricanes (triangles), and intense hurricanes (diamonds) as defined by NHC.

b. Relationship with tropical cyclones

Since it has often been assumed that AEW activity does not vary significantly, it has also been assumed that AEW variability plays little role in tropical cyclone variability. In order to test this we therefore examine whether the AEW variability diagnosed here is related to tropical cyclone variability. Because the time series is short we choose to do this by visual inspection of the time series of the 850-mb AEW activity for box 2 together with the number of named storms, hurricanes, and intense hurricanes for each year (Fig. 13). We choose this box because it is able to characterize the variability in AEW activity leaving the West African coast and we are sure that the box contains no tropical storms (as defined by NHC).

What is striking about these time series is the correspondence between the AEW activity and the tropical cyclone activity for much of the time series. Although before about 1985 there appears to be a negative correlation between AEW activity at 850 mb and tropical cyclone activity, from 1985 onward there is a strong positive correlation, especially for the post-ECMWF reanalysis (post-ERA) period from 1994 to 1998. While of course an objective analysis must consider the whole time series, which is already short, the striking positive correlation seen after 1985 and especially the amazing correspondence for the post-ERA period is certainly worth considering. This analysis offers, for the first time, supportive evidence for the hypothesis that AEW variability is indeed related to Atlantic tropical cyclone variability. As discussed in the introduction, we should not be surprised by this since it is well known that finite-amplitude precursors are known to be required for tropical cyclogenesis (e.g., Gray 1979; Emanuel 1989).

Of course, the relationship is not strong enough to

explain all the variability, because the large-scale environment in the tropical cyclone “main development region” (MDR) must also play a role (cf. Goldenberg and Shapiro 1996). Indeed, it is plausible that variability in large-scale vertical shear and/or SSTs in the MDR could on occasion result in an environment so unfavorable for tropical cyclogenesis that the AEW activity is irrelevant. For example, the fact that 1982 was an inactive tropical cyclone year, despite being an active AEW year, may have been due to anomalously strong vertical shear in the MDR (cf. Thorncroft and Pytharoulis 1999, manuscript submitted to *Wea. Forecasting*).

Further analysis of longer time series is required in order to address these ideas. This should involve analysis of longer reanalysis datasets when they become available as well as continued analysis of subsequent years.

Last, because this work has been at least partly stimulated by the known positive correlation between West Sahel rainfall variability and tropical cyclone variability (Landsea et al. 1998) we briefly consider if the AEW activity diagnosed here has a simple relationship with the West Sahel rainfall variability. The linear correlation coefficient between the West Sahel rainfall index produced by Landsea (see Landsea et al. 1998) and the AEW activity at 850 mb in box 2 is just 0.33, which is only significant at the 85% level. This suggests that the AEW variability diagnosed here does not have a simple relationship with West Sahel rainfall variability and therefore cannot alone account for the correlations presented by Landsea and Gray (1992), for example. Of particular interest is the peak in AEW activity diagnosed in 1995, which was a relatively dry West Sahel year. Future work will consider more closely the relationship between AEW activity and rainfall in other regions of West Africa, for example, in the vicinity of the climatological rainfall maximum at the coast. Alongside this the variability of the AEJ instability must also be considered, which may have some variability independent of West African rainfall variability (cf. Thorncroft and Rowell 1998).

6. Summary and conclusions

Automatic tracking of vorticity centers in ECMWF analyses has been used to develop a 20-yr climatology of African easterly wave activity. Storm track statistics at 600 and 850 mb confirm the complicated easterly wave structures present over the African continent. The rainy zone equatorward of 15°N is dominated by 600-mb activity, and the much drier Saharan region poleward of 15°N is dominated by stronger 850-mb activity, consistent with the recent analysis of Pytharoulis and Thorncroft (1999). Over the Atlantic Ocean there is just one storm track with the 600- and 850-mb wave activity collocated. Based on the growth/decay and genesis statistics, it is concluded that the 850-mb waves, poleward of 15°N over land, generally do not get involved with

the equatorward storm track over the ocean. Instead, there appears to be significant development of 850-mb activity at the West African coast in the rainy zone around (10°N, 10°W), which we propose is associated with latent heat release there.

We have shown that, based on ECMWF vorticity tracking statistics for the 1979–98 period, there is marked interannual variability in AEW activity. It is especially marked for the low-level activity leaving the West African coast at about 10°–15°N.

We have also proposed that, since an important ingredient of Atlantic tropical cyclogenesis is the presence of finite-amplitude low-level vorticity anomalies, variability in the low-level developments at the West African coast may have an impact on Atlantic tropical cyclone variability. To examine this hypothesis, the variability of the 850-mb wave activity at the coast was presented. First, the climatological seasonal cycle showed a peak in September consistent with the climatological tropical cyclone activity peak. The interannual variability of the 850-mb wave activity was also compared with the interannual variability of Atlantic tropical cyclone activity. For the period between 1985 and 1998, a notable positive correlation was seen. This suggests that Atlantic tropical cyclone activity may be influenced by the number of AEWs leaving the West African coast, which have significant low-level amplitudes, and not simply by the total number of AEWs. While we should remain cautious about the statistical significance of this we would strongly recommend further work in this area and continued monitoring. It must be recognized though, that the analyses presented here are based on relatively few assimilated observations in comparison with, say, North America or Europe. Despite this, we know from such studies as Reed et al. (1988a) that the ECMWF analyses are able to have a reasonable representation of AEW developments. We cannot be complacent, however, and while there is some data to support the results here, especially at the coast (cf. Fig. 4), there is a need for more detailed observations of AEWs both at the coast and inland. Future work should also consider the impacts of possible improvements to operational analyses with the introduction of 3D and 4D variational assimilation, which began in 1996.

In interpreting the results presented in this paper it must be remembered that the automatic tracking only tracks vorticity centers with a maximum value greater than $0.5 \times 10^{-5} \text{ s}^{-1}$ that last longer than two days and that travel at least 10°. Our attention is therefore focused on the strongest and most coherent systems. This is very relevant for tropical cyclogenesis. We should note however that the activity over the land diagnosed using this approach is much weaker than over the ocean. This might mean that vorticity centers are not ideal for characterizing the wave activity over the land, either because of the weaker vorticity amplitudes or because of the multicentered nature of the waves. Future work will

consider other methods of tracking the waves over the land including the use of filtered data and different fields. It should be recognized that there are many different methods of diagnosing AEW activity and its associated variability. While this paper has focused on the tracking of vorticity anomalies, other measures such as wind variances, OLR anomalies, or smoother dynamical fields such as meridional wind and streamfunction may yield yet more information on the AEW life cycle and should be examined. Some kind of tracking system combining different fields may give the best results, especially in order to get the transition from the land to the ocean better and also for tracking the weakening systems farther across the Atlantic. This was the approach used by Reed et al. (1988b) in a manual analysis. Future work will explore the possibility of an automated approach to this which may also be useful in operational forecasting.

It should be noted that, while we have provided some evidence for marked easterly wave variability, we do not yet know if there are any large-scale features of the West African monsoon that are related to this, although preliminary analysis suggests a weak link with West Sahel rainfall. Future work will consider in more detail whether there is any simple relationship between wave activity and such things as rainfall variability and African easterly jet variability. This type of analysis will help us to develop our understanding of the important scale interactions occurring in the West African monsoon and particularly in relation to AEWs and whether they play an active or passive role in West African rainfall variability.

Acknowledgments. This research was (in part) supported by the EC Environment and Climate Research Programme (Contract ENV4-CT97-0500, Climate and Natural Hazards). We would like to acknowledge ECMWF for the use of their analyses and Chris Landsea and the NHC for providing the West Sahel rainfall index and the best-track data. We would also like to thank Ioannis Pytharoulis for providing us with Fig. 4. We thank John Molinari, Richard Pasch, and one anonymous reviewer for constructive comments on an earlier version of the paper.

REFERENCES

- Albignat, J. P., and R. J. Reed, 1980: The origin of African wave disturbances during Phase III of GATE. *Mon. Wea. Rev.*, **108**, 1827–1839.
- Avila, L. A., and R. J. Pasch, 1992: Atlantic tropical systems of 1991. *Mon. Wea. Rev.*, **120**, 2688–2696.
- Burpee, R. W., 1972: The origin and structure of easterly waves in the lower troposphere of North Africa. *J. Atmos. Sci.*, **29**, 77–90.
- Carlson, T. N., 1969: Some remarks on African disturbances and their progress over the tropical Atlantic. *Mon. Wea. Rev.*, **97**, 716–726.
- Ceron, J. P., and J. F. Gueremy, 1999: Validation of the space–time

- variability of African easterly waves simulated by the CNRM GCM. *J. Climate*, **12**, 2831–2855.
- Emanuel, K. A., 1989: The finite-amplitude nature of tropical cyclogenesis. *J. Atmos. Sci.*, **46**, 3431–3456.
- Fyfe, J. C., 1999: Climate simulations of African easterly waves. *J. Climate*, **12**, 1747–1769.
- Gibson, J. K., P. Kallberg, S. Uppala, A. Hernandez, A. Nomura, and S. Serrano, 1997: ECMWF Reanalysis Project. Report Series 1. 71 pp.
- Goldenberg, S. B., and L. J. Shapiro, 1996: Physical mechanisms for the association of El Niño and West African rainfall with Atlantic major hurricane activity. *J. Climate*, **9**, 1169–1187.
- Gray, W. M., 1979: Hurricanes: Their formation, structure and likely role in the tropical circulation. *Meteorology Over the Tropical Ocean*, D. B. Shaw, Ed., Royal Meteorological Society, 155–218.
- , 1990: Strong association between West African rainfall and US landfall of intense hurricanes. *Science*, **249**, 1251–1256.
- Hodges, K. I., 1995: Feature tracking on the unit sphere. *Mon. Wea. Rev.*, **123**, 3458–3465.
- , 1996: Spherical nonparametric estimators applied to the UGAMP GCM integration for AMIP. *Mon. Wea. Rev.*, **124**, 2914–2932.
- , 1999: Adaptive constraints for feature tracking. *Mon. Wea. Rev.*, **127**, 1362–1373.
- , and C. D. Thorncroft, 1997: Distribution and statistics of African mesoscale convective weather systems based on the ISCCP Meteosat imagery. *Mon. Wea. Rev.*, **125**, 2821–2837.
- Kwon, H. J., and M. Mak, 1990: A study of the structural transformations of the African easterly waves. *J. Atmos. Sci.*, **47**, 277–292.
- Landsea, C. W., and W. M. Gray, 1992: The strong association between western Sahelian monsoon rainfall and intense Atlantic hurricanes. *J. Climate*, **5**, 435–453.
- , G. D. Bell, W. M. Gray, and S. B. Goldenberg, 1998: The extremely active 1995 Atlantic hurricane season: Environmental conditions and verification of seasonal forecasts. *Mon. Wea. Rev.*, **126**, 1174–1193.
- Lau, K.-H., and N.-C. Lau, 1990: Observed structure and propagation characteristics of tropical summertime synoptic-scale disturbances. *Mon. Wea. Rev.*, **118**, 1888–1913.
- Miller, R. L., and L. S. Lindzen, 1992: Organization of rainfall by an unstable jet with an application to African waves. *J. Atmos. Sci.*, **49**, 1523–1540.
- Molinari, J., D. Knight, M. Dickenson, D. Vollaro, and S. Skubis, 1997: Potential vorticity, easterly waves and eastern Pacific intensification. *Mon. Wea. Rev.*, **125**, 2699–2708.
- Mozer, J. B., and J. A. Zehnder, 1996a: Lee vorticity production by large-scale tropical mountain ranges. Part I: Eastern North Pacific tropical cyclogenesis. *J. Atmos. Sci.*, **53**, 521–538.
- , and —, 1996b: Lee vorticity production by large-scale tropical mountain ranges. Part II: A mechanism for the production of African waves. *J. Atmos. Sci.*, **53**, 539–549.
- Pytharoulis, I., 1999: African easterly waves and their transformation into tropical cyclones in the North Atlantic. Ph.D. thesis, University of Reading, 196 pp. [Available from Department of Meteorology, University of Reading, Whiteknights, Reading, Berkshire, RG6 6AH, United Kingdom.]
- , and C. D. Thorncroft, 1999: The low-level structure of African easterly waves in 1995. *Mon. Wea. Rev.*, **127**, 2266–2280.
- Reed, R. J., 1988: On understanding the meteorological causes of Sahelian drought. *Persistent Meteo-Oceanographic Anomalies and Teleconnections*, C. Chagas and G. Puppi, Eds., Pontificae Academiae Scientiarum, 179–213.
- , D. C. Norquist, and E. E. Recker, 1977: The structure and properties of African wave disturbances as observed during Phase III of GATE. *Mon. Wea. Rev.*, **105**, 317–333.
- , E. Klinker, and A. Hollingsworth, 1988a: The structure and characteristics of African easterly wave disturbances as determined from the ECMWF Operational Analysis/Forecast System. *Meteor. Atmos. Phys.*, **38**, 22–33.
- , A. Hollingsworth, W. A. Heckley, and F. Delsole, 1988b: An evaluation of the performance of the ECMWF operational system in analyzing and forecasting easterly wave disturbances over African and the tropical Atlantic. *Mon. Wea. Rev.*, **116**, 824–865.
- Rowell, D. P., C. K. Folland, K. Maskell, and M. N. Ward, 1995: Variability of summer rainfall over tropical north Africa (1906–92): Observations and modelling. *Quart. J. Roy. Meteor. Soc.*, **121**, 669–704.
- Schubert, W. H., J. J. Hack, P. L. Silva Dias, and S. R. Fulton, 1980: Geostrophic adjustment in an axisymmetric vortex. *J. Atmos. Sci.*, **37**, 1464–1484.
- , P. E. Ciesielski, D. E. Stevens, and H.-C. Kuo, 1991: Potential vorticity modelling of the ITCZ and the Hadley circulation. *J. Atmos. Sci.*, **48**, 1493–1500.
- Thorncroft, C. D., 1995: An idealized study of African easterly waves. Part III: More realistic basic states. *Quart. J. Roy. Meteor. Soc.*, **121**, 1589–1614.
- , and B. J. Hoskins, 1994: An idealized study of African easterly waves. Part II: A nonlinear view. *Quart. J. Roy. Meteor. Soc.*, **120**, 983–1015.
- , and D. P. Rowell, 1998: Interannual variability of African wave activity in a general circulation model. *Int. J. Climatol.*, **18**, 1305–1323.
- , and M. Blackburn, 1999: Maintenance of the African easterly jet. *Quart. J. Roy. Meteor. Soc.*, **125**, 763–786.