Observations of Seven African Easterly Waves in the East Atlantic during 2006

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ABSTRACT

The African Monsoon Multidisciplinary Analyses (AMMA) experiment and its downstream NASA extension, NAMMA, provide an unprecedented detailed look at the vertical structure of consecutive African easterly waves. During August and September 2006, seven easterly waves passed through the NAMMA domain: two waves developed into Tropical Cyclones Debby and Helene, two waves did not develop, and three waves were questionable in their role in the development of Ernesto, Florence, and Gordon. NCEP Global Data Assimilation System (GDAS) analyses are used to describe the track of both the vorticity maxima and midlevel wave trough associated with each of the seven easterly waves. Dropsonde data from NAMMA research flights are used to describe the observed wind structure and as a tool to evaluate the accuracy of the GDAS to resolve the structure of the wave. Finally, satellite data are used to identify the relationship between convection and the organization of the wind structure. Results support a necessary distinction between the large-scale easterly wave trough and smaller-scale vorticity centers within the wave. An important wave-to-wave variability is observed: for NAMMA waves, those waves that have a characteristically high-amplitude wave trough and well-defined low-level circulations (well organized) may contain less rainfall, do not necessarily develop, and are well resolved in the analysis, whereas low-amplitude (weakly organized) NAMMA waves may have stronger vorticity centers and large persistent raining areas and may be more likely to develop, but are not well resolved in the analysis.

1. Introduction

African easterly waves (AEWs) are one of the dominant synoptic-scale features in tropical northern Africa and the east Atlantic. After genesis over Africa, AEWs often propagate westward across the Atlantic and Caribbean and into the east Pacific. AEWs have a typical wavelength of 2000–4000 km, typical phase speed of 7–9 m s$^{-1}$, and period of approximately 3–5 days (Reed et al. 1977). The growth of AEWs is traced to a mixed barotropic–baroclinic instability (Charney and Stern 1962). Two circulation tracks are related to this growth (Carlson 1969b; Burpee 1972, 1974; Reed et al. 1977): one is at the level of the African easterly jet (AEJ) (∼600 hPa) at approximately 9°–11°N, related to the sign reversal of the meridional potential vorticity (PV) gradient, and a second is at low levels north of the AEJ (∼18°–20°N) in the presence of the low-level potential temperature gradient between the hot, dry air of the Sahara to the north of the jet and relatively cooler, moist air to the south of the jet (Pytharoulis and Thorncroft 1999). Previous studies (Pytharoulis and Thorncroft 1999; Ross and Krishnamurti 2007) have concluded that the two circulations are not independent; rather, they exist as a single AEW. Over the ocean, Ross and Krishnamurti (2007) (for 850-hPa vorticity maxima) also conclude that one must distinguish between the convergence of the vorticity tracks and the merging of vorticity centers; convergence could lead to two weak waves, while a merger could lead to a stronger wave. They find that convergence of tracks is more common than merging. AEWs are known to modulate rainfall over northern Africa (Burpee 1974; Payne and McGarry 1977; Duvel 1990). In particular, Payne and McGarry (1977) note that the greatest contribution to rainfall is from disturbances in the southern track. Thorncroft and Hodges (2001), Fink et al. (2004), and Chen (2006) characterize the northern track as dry disturbances.

AEWs are known to trigger tropical cyclones (TCs) in the Atlantic and Caribbean (Riehl 1954; Frank 1970; Landsea 1993). Much uncertainty remains, however,
concerning the role of AEWs in tropical cyclogenesis. Why does one AEW undergo tropical cyclogenesis while another does not? The difficulty in tackling this, and other issues related to AEWs, is related to the data sparseness over northern Africa and the east Atlantic. Daily rawinsondes at Dakar, Senegal, and Sal, Cape Verde, are typically the only conventional observational data between the west coast of Africa and the Caribbean.

The first major field campaign in the region, the Global Atmospheric Research Program’s (GARP’s) Atlantic Tropical Experiment (GATE) in 1974, utilized its high temporal and spatial resolution database to explore the interactions between small-scale tropical features and the larger-scale circulations. The most recent field campaign, staged primarily over northern Africa, the international African Monsoon Multidisciplinary Analyses (AMMA) experiment, provides a detailed dataset over northern Africa. AMMA’s primary goal is to advance knowledge of the relationship of the West African monsoon (WAM) with the physical, chemical, and biological processes over northern Africa. AMMA’s downstream extension is the National Aeronautics and Space Administration (NASA) AMMA (NAMMA), staged from Sal, Cape Verde (Fig. 1), during August and September 2006. Zipser et al. (2009) provides an overview of NAMMA. Experiment goals address questions regarding the difference between developing and nondeveloping easterly waves; the composition, distribution, and microphysics of the Saharan air layer (SAL); and the role of the SAL in cyclogenesis. The NAMMA dataset provides a valuable high temporal and spatial resolution dataset in the east Atlantic. Data include dropsonde observations from the NASA DC-8 medium-altitude research aircraft, rawinsondes, and ground observations at Praia, Cape Verde, and Dakar, Senegal, as well as data from the Tropical Ocean and Global Atmosphere (TOGA) radar at Praia and the NASA Polarimetric (NPOL) radar at Dakar. This unique dataset provides a rare opportunity to document the structure of consecutive easterly waves, both developing and nondeveloping, at multiple pressure levels.

In this study, dropsonde data from NAMMA research flights are evaluated to address key questions regarding the separation of two spatial scales in AEWs. We ask whether it is necessary to distinguish between the large-scale wave and the embedded vorticity maxima and, if so, how does the organization at the two scales impact the fate of the wave? In tracking relative vorticity maxima [two levels; vertically averaged between 925–850 hPa and 700–600 hPa in the 40-yr ECMWF Re-Analysis (ERA-40)]. Kerns et al. (2008) conclude that it is important to distinguish between vorticity maxima, which can be short-lived, periodic, and multicentered, and the large-scale, longer-lived, wave. This agrees with Hopsch et al. (2007), who conclude that the two measures of AEW activity—the synoptic-scale wave trough and convectively generated subsynoptic-scale vorticity centers—are not necessarily related. Rather, coherent subsynoptic-scale structures can exist without the presence of a strong AEW, and synoptic-scale AEWs can exist without coherent subsynoptic-scale structures. Likewise, Reed et al. (1988) and Thorncroft and Hodges (2001) caution on using vorticity maxima to characterize wave activity over land in weak or multicentered waves. In this paper, high temporal resolution rawinsonde data at Praia, Cape Verde, and Dakar, Senegal, also aid an evaluation of the utility of the meridional wind reversal (from northerlies to southerlies) for identifying wave passages at point stations.

The scale separation is inherently important in tropical cyclogenesis. Several studies have identified the relationship between vorticity maxima and AEW variability with tropical cyclone activity. Thorncroft and Hodges (2001) suggest that the variability of low-level vorticity maxima embedded within the large-scale easterly wave compares well with tropical cyclone variability. In particular, the maximum track density of 600- and 850-hPa vorticity maxima is co-located with the main development region (MDR) for Atlantic tropical cyclones. Although in some disagreement, Hopsch et al. (2007) conclude that subsynoptic-scale structures correlate poorly with tropical cyclone activity at interannual time scales, whereas synoptic-scale AEWs are well correlated. An important relationship between diabatically generated subsynoptic-scale potential vorticity (PV) anomalies from rainfall (maximum during September and October) over the Guinea Highlands and tropical cyclogenesis has, however, been identified previously (Berry and Thorncroft 2005; Hopsch et al. 2007). Furthermore, Thorncroft and Hodges (2001), Hopsch et al. (2007), and Kerns et al. (2008) note the relative unimportance of the low-level...
northern track vorticity maxima in tropical cyclogenesis. The development of new low-level vorticity maxima in the southern track is identified in these studies as the most relevant disturbance track for tropical cyclogenesis, whether or not preceded by a vorticity maximum within the track. In this study, using vorticity maxima tracks from model analysis in combination with satellite observations, the question of whether the northern disturbance track is important for cyclogenesis downstream (during NAMMA) will be addressed. Likewise, the collection of dropsonde observations, statistics on areas of cold cloud and rainfall from satellite platforms, and model analysis will be used to address how the organization of each wave, as it relates to the synoptic-scale wave trough and subsynoptic-scale vorticity centers, may influence the fate of the system (developing or nondeveloping).

During NAMMA, seven AEWs were observed by the NASA DC-8 and ground-based observing stations. Two waves quickly became tropical cyclones, Debby and Helene, off the coast of Africa; two waves did not develop; while three waves seem to have played some role in the development of TCs Ernesto, Florence, and Gordon. The focus of this paper is to use the NAMMA dataset to explore, quantitatively, the characteristics of these seven AEWs. For each wave, we

1) use the National Centers for Environmental Prediction (NCEP) Global Data Assimilation System (GDAS) operational analysis to describe the track of, and relationship between, the synoptic-scale wave trough and subsynoptic-scale vorticity maxima;
2) use dropsonde and rawinsonde data to describe the actual wind structure of the waves, which also facilitates evaluation of the GDAS analysis; and
3) use satellite data to identify the relationship between convection and the scale and structure (organization) of wave features.

Of particular interest is the information that dropsonde data provide on the wave structure, which the GDAS analysis does not replicate well.

The paper is organized as follows: Section 2 provides a description of the NAMMA dataset used in the study, including satellite data, and the tracking methods employed. Section 3 reviews each of the seven waves individually using a combination of the GDAS analysis and dropsonde data from the DC-8. Section 4 presents the rainfall and cold cloud statistics for the waves. Section 5 discusses the results from tracking vorticity maxima and wave troughs in the GDAS analysis and relates results of the dropsonde–GDAS analysis comparison with rainfall and cold cloud statistics. Finally, section 6 presents a summary and conclusions.

2. Data and methodology

a. NAMMA experiment data

Data from NAMMA in this study includes 197 dropsondes from 13 NASA DC-8 missions, all based from Sâl, Cape Verde. Obtained at half-second intervals, measurements include pressure, dry-bulb temperature, dewpoint temperature, relative humidity, zonal (u) and meridional (v) wind components, wind speed, wind direction, latitude, longitude, and GPS altitude.

A special high frequency set of rawinsonde measurements from Praia, Cape Verde, began on 18 August and ended on 14 September with five to seven launches daily. Rawinsonde measurements over northern Africa include observations from Dakar, Senegal (four times daily), as well as Bamako, Niamey, Ouagadougou, Tombouctou, and Tambacounda (Fig. 1).

b. Satellite data

To describe the convective characteristics of the waves, data from a number of remote sensing platforms are obtained. Cold cloud statistics are computed from infrared (IR) brightness temperature (Tb) data available from the National Climatic Data Center (NCDC) Hurricane Satellite (HURSAT) North Atlantic basin dataset, derived from International Satellite Cloud Climatology Project (ISCCP) B1 data (Knapp and Kossin 2007). The dataset consists of merged Geostationary Operational Environmental Satellite (GOES) and Meteosat Tb data, available 3 hourly and gridded at 8-km horizontal resolution. Also, 36.5- (14 × 8 km² footprint) and 89-GHz (6 × 4 km² footprint) microwave Tb data is from the Advanced Microwave Scanning Radiometer–Earth Observing System (AMSR-E). The Tropical Rainfall Measuring Mission (TRMM) Microwave Imager (TMI) provides 37- and 85-GHz brightness temperatures (16 × 9 and 7 × 5 km² footprints, respectively). The TRMM Precipitation Radar (PR) provides vertical profiles of reflectivity (4.3 × 4.3 km² footprint), while TRMM’s Visible and Infrared Scanner (VIRS) provides IR brightness temperature (2 × 2 km² footprint). Finally, 3-hourly 0.25° × 0.25° gridded rain rates are available from the level-3 TRMM algorithm, 3B42. The algorithm adjusts rain rates derived from geosynchronous IR observations with TRMM estimated surface rain and hydrometeor structure.

c. NCEP GDAS description

The Global Data Assimilation System, run by NCEP and obtained from the University Corporation for Atmospheric Research (UCAR) Data Support Center, analysis is utilized to track 700- and 925-hPa vorticity maxima and the 700-hPa wave trough. The GDAS uses
a T254 Gaussian grid, equivalent to a 1° × 1° latitude/longitude grid spacing with global coverage. The GDAS archive provides fields four times daily (0000, 0600, 1200, and 1800 UTC) for the following variables at 64 vertical pressure levels: temperature, height, absolute vorticity, vertical velocity, relative humidity, and u and v wind. The analyses are produced by using the current analysis to initialize the Global Forecast System (GFS), which then produces 3-, 6-, and 9-h forecasts. These become “pseudo-observations” and are combined with actual observations to create the next analysis field. This technique is described in Derber et al. (1991) and Parrish and Derber (1992) with additional modifications outlined in 2006 NCEP and UCAR technical procedure bulletins. No dropsonde data from NAMMA were assimilated into the GDAS analysis (Steve Lord, NCEP, 2006, personal communication), so they can be used as an independent source of validation data.

d. Tracking methodology

Tracking methods can be categorized as manual, automatic, or statistical. Manual tracking methods (Carlson 1969a,b; Reed et al. 1988; Fink and Reiner 2003; Fink et al. 2004; Chen 2006) include objectively tracked easterly waves using pressure-level streamlines; 700-, 850-, and 925-hPa relative vorticity; cloud clusters; 700-hPa streamfunction (Berry et al. 2007); and/or meridional wind fields. Automatic tracking methods (Thorncroft and Hodges 2001; Hopsch et al. 2007) use similar variable fields but are constrained by selection criteria or thresholds. Statistical methods (Burpee 1972, 1974; Albignat and Reed 1980; Reed et al. 1988; Duvel 1990; Lau and Lau 1990; Diedhiou et al. 1999; Fyfe 1999; Pytharoulis and Thorncroft 1999) often involve tracking using (3–4 day) bandpass filtered or composited vorticity and/or meridional wind. The reader is referred to Kerns et al. (2008) for a more extensive review of tracking methodology for easterly waves.

A manual tracking methodology is employed in this study. Since one must distinguish between the large-scale wave trough and embedded vorticity maxima (Hopsch et al. 2007; Kerns et al. 2008), both the wave trough (defined as the zero meridional wind contour that separates the northerlies west of the wave trough from southerlies to the east) and the vorticity maxima are tracked independently in this study. Using the GDAS absolute vorticity analyses, absolute vorticity maxima (henceforth, “vorticity maxima”) following the northern track at 925 hPa and following the southern track at 700 hPa and exceeding a threshold of $7 \times 10^{-5}$ s$^{-1}$ (slightly larger than the background value of the Coriolis parameter) are tracked. Vorticity maxima are first tracked when the feature is seen consistently in the analysis for at least two days and ends when the vorticity maximum drops below the threshold, is north of 30°N or west of 60°W, or cannot be identified in the following analysis time (no longer consistent). The 700-hPa GDAS meridional wind fields are utilized to track the wave trough. Wave trough tracking begins at the same start time of the vorticity tracking and ends when the zero meridional wind contour is no longer apparent in the analysis field. All GDAS analysis fields are raw with no filtering or smoothing applied.

3. Overview of the seven NAMMA AEWs

Between 19 August and 12 September 2006, seven easterly waves passed through the NAMMA domain. For each wave, there are dropsonde and rawinsonde observations available, which provide an unusual capability for evaluating the quality of the GDAS analysis of the waves as they passed through the east Atlantic. Figures 2a and 2b show a summary of GDAS-analyzed vorticity maxima tracks at 925 and 700 hPa. Consistent with climatologies stated in literature, five of the 925-hPa vorticity maxima tracks are north of the 700-hPa tracks. Wave 2 (developing) and wave 3 (nondeveloping) are exceptions. Over land, the mean latitude of the 925-hPa track is approximately 17°N and for 700 hPa, at 13°N. Over the ocean, the mean latitude of both 925- and 700-hPa vorticity maxima tracks is approximately 15°N. Over land, vorticity maxima often have very short lifetimes; thus, tracking vorticity maxima in the GDAS analysis back into eastern Africa is restricted by the
consistency criteria outlined in the methodology: only one vorticity maximum (associated with wave 7, developing) is tracked well east of the Greenwich meridian. The two recurving tracks over the ocean are TC Debby (wave 2) and TC Helene (wave 7).

Rawinsonde time series of relative humidity, zonal (u) wind, meridional (v) wind, and temperature anomaly from Dakar (Fig. 3, 6-hourly) and Praia (Fig. 4, 4-hourly) provide a more detailed look at the observed wave passages. There are only two distinct wave passages in the observations at Dakar at 700 hPa (where the v wind goes from green and blue, northerlies, to yellow and orange, southerlies): one associated with nondeveloping wave 3 on day 237 (25 August) and the other with developing wave 7 (Helene) on day 254 (11 September). Distinct wave passages at Praia in the observations include nondeveloping wave 3 on day 238 (26 August) and the two developing waves, wave 2 (Debby) on day 234 (22 August) and wave 7 on day 256 (13 September). The AEJ (zonal wind) exceeds 20 m s$^{-1}$ during several periods at Dakar and Praia: during the approximate period of wave-1 passage, between waves 2 and 3, between the passage of waves 4 and 5, and during wave-7 passage. For those waves, 700-hPa vorticity maxima generally pass to south of the rawinsonde sites. Evaluating both the RH and temperature anomaly data reveals noticeable SAL outbreaks: preceding wave-3 passage at Dakar there is low RH at low to mid levels (900–550 hPa) with a temperature anomaly of up to 4°C (at 900–800 hPa). At Praia, however, the SAL not only precedes wave 3 but is also prevalent following the passage. At both rawinsonde sites, the passage of the 925-hPa vorticity maximum associated with wave 1 is dominated by the SAL, with up to a 5°C–6°C warm anomaly at 850 hPa at Praia. Perhaps significantly, there are periods of low RH and a warm anomaly preceding the passage of both developing waves (2 and 7) at Praia. Since the centers of both developed storms pass south of Praia, one may anticipate that the SAL was prevalent to the north of the centers in both waves 2 and 7.
The temperature anomaly also provides an opportunity for a comparison to the composite anomalies presented in Reed et al. (1977) for vorticity centers associated with eight GATE AEWs. Of all of the waves in NAMMA, wave 3 most mirrors the composite of Reed et al. (1977). Both the 925- and 700-hPa vorticity centers pass directly over Praia and Dakar. The temperature anomaly at Dakar (Fig. 3) reveals a warm anomaly ($\sim 3^\circC$) in the northerlies between 900–600 hPa and a cold anomaly ($\sim -3^\circC$) between 550 and 450 hPa in the northerlies preceding passage. In the wave trough and after passage, there is a cold anomaly ($\sim -2^\circC$) from the surface to 700 hPa and a warm anomaly ($\sim -2^\circC$) above, from 600 to 300 hPa. The wave-3 passage at Praia (Fig. 4) is similar; however, the depth and magnitude of the low-level cold anomaly during and after passage is shallower (well below 700 hPa, indicated in the composite) and colder than at Dakar. An important difference from the composite is that the magnitudes of the anomalies are much greater in the individual NAMMA waves. Compositing is a smoothing process, and bandpass filtering for 2–6 days performed by Reed et al. (1977) may contribute to the difference in magnitude.

Other waves present unique differences from the composite. At Dakar, the wave-6 passage is primarily cold core through midlevels, although immediately following passage at Praia the warm anomaly at 600–500 hPa is more apparent. At both Dakar and Praia, for waves 4 and 5 there is a cold anomaly from 850–700 hPa, while a warm anomaly is present below 925 hPa. Overall, the high-resolution rawinsonde time series illustrates waves with complex wind and thermal structures, which vary in part owing to the location of the site relative to the wave and in part because AEWs vary themselves. In the following subsections, a brief overview of each NAMMA wave is provided in the context of the GDAS analysis and snapshots from dropsonde observations.

### a. Wave 1

#### 1) GDAS

For all waves, the GDAS analysis absolute vorticity and meridional wind fields are the primary tools for tracking and describing the structure of the vorticity maxima and wave trough. GDAS-analyzed vorticity maxima and wave trough tracks for the first wave of
NAMMA (Fig. 5) indicate that the wave exited the coast on 18 August. The latitudes of both the 925- and 700-hPa vorticity maxima tracks are consistent with the typical tracks in the literature.

An important aspect of wave 1 is that, over the ocean, the GDAS-analyzed 700-hPa wave trough begins to outrun the 700-hPa vorticity maximum on 19 August (the difference is as large as 10° longitude by 1200 UTC 21 August) and becomes increasingly tilted in the northwest–southeast direction. The 700-hPa vorticity maximum can no longer be tracked after 21 August. At 925 hPa, the GDAS-analyzed vorticity maximum in the northern track (Fig. 5) dissipates quickly after departing the coast as it is characteristically dry (lacking rainfall) owing to the overwhelming presence of the SAL (Fig. 6). An important vorticity maximum and closed circulation at 925 hPa (not included in the original tracking; Fig. 5) appears in the analysis over the ocean in the southern track on 20 August and appears to have developed between 19 and 20 August. Figure 7 provides a closer look at the GDAS analysis wind vectors at 1200 UTC on 19 (Fig. 7a) and 20 August (Fig. 7b), which clearly depict a tight circulation along 8°N at 925 hPa. In contrast to the dry, low-level northern track vorticity maximum tracked from northern Africa (Fig. 5), the circulation at 925 hPa depicted in Fig. 7b is characterized by widespread ITCZ convection (Fig. 6).

The ambiguities in the GDAS analysis vorticity maxima and wave trough tracks are relevant: on 24 August, the National Hurricane Center (NHC) declared a depression (eventually TS Ernesto) at 1800 UTC near 12.7°N, 61.6°W. We may speculate that a westward movement of the 20 August vorticity maximum at 6° per day would place it near this genesis location. But, the connection between the wave trough and vorticity maxima with the cyclogenesis event is not clear, so the question remains: What is the contribution of the lagging elongated vorticity maxima at midlevels and remaining vorticity maxima in the ITCZ at low levels?

2) DROPSONDE OBSERVATIONS

Dropsonde observations from research flights provide unique snapshots detailing aspects of the actual AEW vertical wind structure and are utilized to help evaluate the GDAS analysis. All dropsonde data shown are time–space adjusted to 1200 UTC using the vorticity maxima propagation speed. On 19 and 20 August (Figs. 7a and 7b, respectively) dropsonde wind data (barbs) at 700 hPa show very little evidence of wave trough curvature;

![Fig. 6. Visible satellite imagery from 1145 UTC 20 Aug 2006 with locations (open circles) of the vorticity maximum at 925 hPa tracked from northern Africa and the vorticity maximum that appears near the ITCZ (not tracked in Fig. 5). The SAL is identified by the white haze with its approximate border indicated by the thick black line.](image_url)
however, if the trough is located where the analysis indicates, the dropsondes are to the east and the research flights did not sample the wave trough. Likewise, on 19 August the GDAS analysis compares least favorably to 925-hPa dropsonde data at latitudes south of 10°N and is most robust near the AEJ (~15°N) at 700 hPa. On 20 August (Fig. 7b), cyclonic curvature in the 925-hPa dropsonde wind data seemingly indicates that the northeast quadrant of a circulation is sampled during the flight, which supports the tight circulation in the GDAS analysis at the same level. To quantify the differences, the root-mean-square error (RMSE) is computed for each comparison figure for the dropsonde value and the nearest GDAS analysis grid point. In Fig. 7a the RMSE for the $u$ wind is 2.9 (2.8) m s$^{-1}$ at 925 (700) hPa and for the $v$ wind, 3.7 (4.8) m s$^{-1}$ at 925 (700) hPa. The $u$-wind RMSE, Fig. 7b, is 2.5 (4.4) m s$^{-1}$ at 925 (700) hPa, and the $v$-wind RMSE 5.2 (2.4) m s$^{-1}$ at 925 (700) hPa. One must use caution in the interpretation of the RMSE due to the high sensitivity to large errors; rather, one should combine the RMSE with the visual inspection to completely describe the differences. For example, a phase shift of the wave between the observed and GDAS analysis wave may cause a high RMSE but visually look closely comparable. Overall, the dropsonde data provide valuable information about the observed wave structure; the observations indicate that wave 1 is a low-amplitude ("weak") wave that is initially disorganized after leaving Africa. That is, a coherent high-amplitude midlevel wave trough and distinct low-level circulations, which are typical of composite easterly waves described in literature, are lacking. The GDAS analysis is not robust in this case (the exception is the tight circulation that may be present at 925 hPa on 20 August). A partial explanation for the differences between the GDAS analysis and the observed wind structure may be the effect of widespread ITCZ convection (Fig. 6). Meso-scale processes may complicate and otherwise mask synoptic-scale easterly wave features; the analysis at 1° resolution does not seem to do well in this scenario.

b. Wave 2

1) GDAS

Wave 2 exited the coast of Africa on 21 August and quickly became TS Debby, the first of two unambiguously
developing waves during NAMMA. The GDAS-analyzed 700- and 925-hPa vorticity maxima and 700-hPa wave trough tracks are provided in Fig. 8. The consistency criteria were not met for either the wave or vorticity maxima until 20 August, the last day the system was over land. In fact, a consistent low-level vorticity maximum is not prevalent until organized along the Guinea coast at 10°N on 20 August. Soon after moving over ocean, the vorticity maxima quickly became vertical and intensified, and Debby formed on 21 August.

2) CYCLOGENESIS

A number of authors (Thorncroft and Hodges 2001; Berry and Thorncroft 2005; Hopsch et al. 2007) have noted genesis of vorticity maxima near the Guinea Highlands that they ascribe to latent heating from convection in the monsoon trough and relate such vorticity maxima to subsequent cyclogenesis downstream over the east Atlantic. After analyzing IR and microwave satellite imagery of the convection during the hours prior to classification, this type of vorticity genesis event may help to explain the formation of Debby. Intense mesoscale convective systems (MCSs) along the coast of West Africa are easily seen in a couple of fortuitous overpasses from TRMM on 21 August. The first overpass, approximately 16 h prior to classification (Fig. 9), shows two noticeably intense MCSs. The evidence for intense convection is the depressed 85-GHz polarization-corrected temperature (PCT), indicating significant ice scattering (Spencer et al. 1989; Mohr and Zipser 1996; Zipser et al. 2006). The northernmost MCS is a squall line just south of Dakar and the second is an MCS originating off the coast of Guinea. The intensity (based on minimum 85-GHz PCT) of the MCSs is within the strongest 1% of TRMM precipitation features (defined in Liu et al. 2008) within 5° of vorticity maxima tracks in

the NAMMA focus region for June–September (JJAS) 1998–2007. An overpass 10 h later (Fig. 10, 6 h prior to classification) contains MCSs of slightly less intensity, but still in the top 1% of the regional climatology, and much greater organization.

3) DROPSONDE OBSERVATIONS

The research flight into Debby on 23 August provides rare observations of a tropical storm in the east Atlantic. Dropsonde wind data (Fig. 11) reveal a well-developed, tight circulation at low and mid levels with a wind maximum of over 30 m s$^{-1}$ at 600–700 hPa. In situ data from the flight indicated a 5°–6°C temperature associated with a warm core at 700 hPa (center temperature of 14°C). [See Zipser et al. (2009) for a more extensive observational look at Debby.] In Fig. 11 the RMSE for $u$ wind is 8.1 (9.2) m s$^{-1}$ at 925 (700) hPa and
for \(v\) wind 6.3 (6.2) m s\(^{-1}\) at 925 (700) hPa. Comparing the dropsonde wind data to GDAS analysis indicates an important deficiency in the analysis; at 700 hPa, the GDAS indicates an open-wave wind structure rather than the tight circulation actually present. Comparing the core temperature of the GDAS analysis “cyclone” (280 K), the analysis is 7 K cooler at 700 hPa than observed (287 K) and is cold core rather than warm core—a fundamental problem with the GDAS analysis, although it is not expected that its \(1^\circ\) resolution could give an accurate description of such a small tropical cyclone.

c. Wave 3

1) GDAS

The clearest example of a nondeveloping wave from NAMMA is wave 3. Exiting the coast of Africa on 25 August, wave 3 exhibits a high-amplitude midlevel wave trough and distinct low-level circulations. Both the GDAS-analyzed 925- and 700-hPa vorticity maxima (Fig. 12) emerge out of the Sahara and are very easily tracked throughout the duration of the wave. The 700-hPa vorticity maximum track is, however, not characteristic of climatology; the track is much farther north. While the magnitude of the 925-hPa vorticity maximum (not shown) undergoes a diurnal maximum (during the day) and minimum (during night) due to the strong sensible heat flux from the Sahara over land, both the 925- and 700-hPa vorticity maxima weaken as they track across the Atlantic.

2) DROPSONDE OBSERVATIONS

Dropsonde wind data on 25 (Fig. 13) and 26 (not shown) August validate the high-amplitude character of wave 3 seen in the GDAS analysis. Unlike previous flights into waves 1 and 2, the GDAS analysis compares very closely with the dropsonde wind data; thus, we have high confidence in the GDAS analysis in this case. In Fig. 13 the RMSE for the \(u\) wind is 3.5 (2.3) m s\(^{-1}\) at 925 (700 hPa) and for the \(v\) wind 2.0 (4.1) m s\(^{-1}\) at 925 (700) hPa. Because the 700-hPa wave trough and vorticity maximum all exhibited excellent continuity for more than 10 days over land and ocean, we suggest that the relative ease of tracking is linked to the lack of large persistent raining areas or MCSs, which would otherwise drive intense local circulations that could mask the wave features. The convection has a characteristic diurnal variation over land and exhibits a peak in convection on the coast due to strong northward moisture advection from lower latitudes, but becomes less prevalent over the ocean once the SAL spreads throughout the circulation (Praia rawinsonde data, Fig. 4 on days 237–239). Despite the impressive high-amplitude character of wave 3, it does not develop.

d. Wave 4

1) GDAS

The fourth wave of NAMMA exited the coast on 1 September. Like wave 1, the GDAS-analyzed wave trough outruns the vorticity maxima over the central
Atlantic, leading to a lagging of the 700-hPa vorticity maximum behind the wave trough beginning on 2 September (Fig. 14). By 0000 UTC 3 September, the GDAS-analyzed 700-hPa vorticity maximum becomes elongated and is no longer consistently tracked. As a result, the link between the GDAS-analyzed features of wave 4 and the genesis of TC Florence (NHC declared a tropical depression at 1800 UTC on 3 September at 14.1°N, 39.4°W) is unclear. As in wave 1, the 925-hPa northern track vorticity maximum weakens shortly after leaving the coast and has no obvious relevance to downstream cyclogenesis.

2) DROPSONDE OBSERVATIONS

Dropsonde data from a flight on 1 September (Fig. 15) provides detailed observations of the actual wave structure when wave 4 is just off the coast of Africa. Like wave 1 (and in stark contrast with wave 3), the dropsonde wind data show a disorganized wind structure; no distinct synoptic-scale circulations are present at either low or mid levels, and observations provide little support for a northwest–southeast tilted wave trough analyzed at 700 hPa along 23°W in the GDAS analysis. At 700 hPa the dropsondes indicate that an east–west oriented vorticity maximum is located along 11°N, but this feature is totally absent in GDAS. There is room for speculation that this cyclonic region, associated with ITCZ convection, may have moved westward to a position close to the genesis location of Florence. In Fig. 15 the RMSE for the $u$ wind is 3.1 (4.4) m s$^{-1}$ at 925 (700) hPa and for the $v$ wind 3.3 (4.2) m s$^{-1}$ at 925 (700) hPa. Similar to wave 1, the complex wind field indicated by the dropsonde data may be a result of subsynoptic-scale processes, such as organized deep convection in and near the ITCZ (visible satellite imagery, Fig. 16), that mask the larger-scale wave features; the GDAS analysis does not compare favorably with the observed wave structure in this scenario.

e. Wave 5

1) GDAS

Exiting the west coast of Africa on 2 September, the GDAS-analyzed 925- and 700-hPa vorticity maxima are
tracked (Fig. 17) along 20° and 15°N, respectively, converge off the coast of Africa, and then tracked until 7 and 8 September at which time the maxima undergo elongation due to deformation east of TC Florence. Since Gordon is not classified until 10 September (the NHC declared a tropical depression at 1800 UTC at 20.2°N, 53.8°W), no conclusion is drawn as to whether the vorticity maxima tracked in the GDAS analysis associated with wave 5 played a role in the genesis of Gordon.

2) DROPSONDE OBSERVATIONS

Research flights on 3 (Fig. 18a) and 4 (Fig. 18b) September provide a detailed look at the low and midlevel wind structure on consecutive days. On both days, dropsonde wind data show a high-amplitude, northeast–southwest tilted wave trough at 700 hPa and support the presence of a well-defined, synoptic-scale circulation at low levels. Like wave 3, similar features in the GDAS analysis are well supported by the observations. In Fig. 18a the RMSE for the $u$ wind is 2.2 (3.6) m s$^{-1}$ at 925 (700) hPa and for the $v$ wind 2.2 (3.9) m s$^{-1}$ at 925 (700) hPa. In Fig. 18b the RMSE for the $u$ wind is greater, 3.9 (4.8) m s$^{-1}$ at 925 (700) hPa, and for the $v$ wind, 2.5 (2.5) m s$^{-1}$ at 925 (700) hPa. Although development is questionable over the central Atlantic, the total rainfall characteristics of the wave compare closely to rainfall characteristics of the nondeveloping waves of NAMMA (see next section).

f. Wave 6

1) GDAS

In the GDAS analysis, no consistent 925-hPa vorticity maxima are tracked with wave 6 (Fig. 19); however, a 700-hPa vorticity maximum is tracked from 10° to 40°W, leaving Africa on 8 September. The GDAS-analyzed 700-hPa vorticity maximum tracks coincidentally with the analyzed 700-hPa wave trough for much of its life over the east Atlantic until the magnitude of absolute vorticity drops below the threshold ($7 \times 10^{-5}$ s$^{-1}$) at 37°W; a weak wave trough is tracked consistently westward until 14 September.

2) DROPSONDE OBSERVATIONS

Dropsondes from the two dedicated flights [8 September (Fig. 20) and 9 September (not shown)] show considerable

![Fig. 14. As in Fig. 5 but for wave 4.](Image)

![Fig. 15. As in Fig. 7 but for the flight into wave 4 on 1 Sep 2006.](Image)
agreement with the GDAS analysis of a low-level circulation and high-amplitude wave trough at 700 hPa. In Fig. 20 the RMSE for the $u$ wind is $2.3 (3.6)$ m s$^{-1}$ at 925 (700) hPa and for the $v$ wind $1.7 (4.2)$ m s$^{-1}$ at 925 (700) hPa. Wave 6 shares common characteristics with wave 3; that is, the wave has large amplitude, little rainfall (see the next section), and does not develop.

g. Wave 7

1) GDAS

The second of two unambiguously developing waves is the seventh and final wave of NAMMA; wave 7 became Hurricane Helene. In contrast to the other developing case (wave 2), wave 7 is consistently tracked in the GDAS analysis from eastern Africa (Fig. 21) through genesis off the coast of Africa. In fact, the vorticity maxima and wave trough tracks in the GDAS analysis are the longest and most consistent of any of the NAMMA waves. Furthermore, there appears to be a clear merging of the wave, the northern track low-level and southern track midlevel vorticity maxima just off the coast.

2) DROPSONDE OBSERVATIONS

The flight into developing Helene, on 12 September, was the final research flight of NAMMA. In contrast to Debby on 23 August, the circulation center on 12 September is cool and has weaker inner core winds [Zipser et al. (2009) provides a more detailed look at flight observations of the depression on 12 September]. Though few in number, the dropsonde wind data (Fig. 22) and in situ data at multiple levels are in fairly good agreement with GDAS, indicating that this larger-scale circulation is resolvable even at a 1° analysis grid. In Fig. 22 the RMSE for the $u$ wind is $6.5 (8.0)$ m s$^{-1}$ at 925 (700) hPa and for the $v$ wind $6.9 (5.1)$ m s$^{-1}$ at 925 (700) hPa.

4. Cold cloud and rainfall statistics

Rainfall and cold cloud statistics of each vorticity maximum tracked are compared in this section. Table 1 gives statistics representing averages (over the duration of the vorticity maxima) of TRMM 3B42 data within a $10^\circ \times 10^\circ$ box centered on each 3-hourly 700- and 925-hPa vorticity maximum. Statistics include the fraction of raining pixels (of total pixels within the box) and fraction of the raining pixels with an intense rain rate ($\geq5$ mm h$^{-1}$). Similarly, Table 2 displays statistics using the ISCCP B1 3-hourly IR $T_b$ data: statistics include the fraction of total pixels within the box that have a $T_b \leq 235$ and 210 K, as well as the mean and mean minimum $T_b$ of all the pixels within the box over the duration of each vorticity maximum.

In general, given substantially lower fractions of raining pixels, the (northern track) 925-hPa vorticity maxima

![Fig. 16. Visible satellite imagery at 1145 UTC 1 Sep 2006 with the locations (open circles) of the 925- and 700-hPa GDAS-analyzed vorticity maxima.](image-url)

![Fig. 17. As in Fig. 5 but for wave 5.](image-url)
are less rainy than the (southern track) 700-hPa vorticity maxima. The fraction of raining pixels that have intense rain rates is, however, often greater for 925-hPa maxima. The 700-hPa vorticity maxima associated with non-developing waves (3 and 6) exhibit the lowest fraction of raining pixels, consistent with the hypothesis that the presence of the SAL inhibits deep convection and widespread rainfall; not surprisingly, vorticity maxima associated with developing waves (2 and 7) exhibit the largest fractions. Furthermore, wave 2 (Debby) exhibits much less total rainfall than wave 7 (Helene). One may, however, question the utility of the fraction of raining pixels statistic to distinguish developing vorticity maxima from nondeveloping since those that do not immediately develop (waves 1 and 4) had fractions comparable to the two developing cases. In fact, the

FIG. 18. As in Fig. 7 but for (a) the first flight into wave 5 on 3 Sep 2006 and (b) the second flight on 4 Sep 2006.

Fig. 19. As in Fig. 5 but for wave 6.
700-hPa vorticity maximum associated with wave 1, which is primarily ITCZ convection, has a greater fraction of intense rain rate than the developing cases. Wave 5, which does not immediately develop, has rainfall statistics comparatively closer to those of vorticity maxima tracked with nondeveloping waves 3 and 6.

While the rainfall statistics show some clear differences between the waves, the IR statistics in Table 2 are much less discriminating. They indicate an interesting result in which the mean minimum $T_b$ for vorticity maxima developing immediately off the coast (waves 2 and 7) are comparable to those that do not immediately develop (waves 1, 4, and 5); therefore, the mean minimum $T_b$ for NAMMA waves does not distinguish developing from nondeveloping situations well. Only vorticity maxima for nondeveloping waves 3 and 6 have a low fraction of cold cloud ($\approx 235$ and 210 K) and mean IR $T_b$ (wave 3 has a mean IR $T_b$, 26 K greater than wave 7). Although the overall statistics tend to show little distinction between waves that develop off the coast and those that do not, one must be cautious in applying these results as a more general conclusion. In line with the result of lower fractions of raining pixels, the (northern track) 925-hPa vorticity maxima have much less fractions of cold cloud than vorticity maxima at lower latitudes.

5. Discussion

The evidence from the seven waves of NAMMA supports the conclusions of Kerns et al. (2008) that the vorticity maxima within the large-scale easterly wave trough are not “obstacles” to tracking. Rather, vorticity maxima must be distinguished from the wave trough. Over the ocean, the GDAS-analyzed 700-hPa wave troughs for waves 1 and 4 become more northwest–southeast tilted with time and outrun both the low- and midlevel vorticity maxima. Those lagging vorticity maxima, which in both waves are located near the ITCZ, become increasingly elongated (oriented northwest–southeast) and are no longer tracked; perhaps significantly, cyclogenesis occurs shortly thereafter on the southeast edge of the vorticity maxima in the GDAS analysis. Consequently, we classify both waves 1 and 4 as possibly contributing to cyclogenesis, although in both waves the connection between the wave/vorticity maxima and cyclogenesis is unclear because the critical events occur (or not) after detailed observations are no longer available.

Over land, identification of the midlevel wave trough and consistent tracking of vorticity maxima in the GDAS analysis is complicated by the characteristically short lifetime of vorticity maxima or the presence of multiple vorticity centers. This is related to the generation of
subsynoptic-scale vorticity maxima from convection (such as midlevel mesoscale convective vortices) at lower latitudes (10°–15°N) and strong sensible heat fluxes at low levels at higher latitudes (15°–20°N). This point has not gone unnoticed; Reed et al. (1988) and Thorncroft and Hodges (2001) have concluded that in weak, or multicentered waves, vorticity maxima are not ideal for characterizing wave activity over land. Multicentered vorticity maxima also complicate tracking in the GDAS analysis over the ocean, in particular near convectively active portions of the ITCZ in the presence of ambient cyclonic shear.

In a related issue, high-frequency soundings at Praia and Dakar (Figs. 3 and 4) indicate that identifying a wave passage (at a station) using the traditional technique of a northerly to southerly wind transition, while sufficient for statistical compositing, may not be sufficient for identifying individual wave passages. Sign reversals of the meridional wind may not only signify a passage of a wave trough but can be mistaken for a squall-line passage or a vorticity maximum passage not directly associated with the AEW. Conversely, the absence of a distinct meridional wind sign reversal does not mean there is no wave passage; if the strongest part of the circulation does not pass near the station then there may be no signal, while a transition from weak southerlies (with strong easterly component) to strong southerlies (or southwesterlies) may represent the passage of a weak or tilted wave axis. For example, wave 1 has no clear passage at low levels at Dakar, whereas at Praia a clear wave passage is absent at midlevels (above 800 hPa). Likewise, at Dakar, at approximately the same time that the GDAS analysis indicates a wave-5 passage, the data show a transition from a strong easterly wind (25 m s\(^{-1}\)) at 700 hPa, with a weak southerly component, 3 m s\(^{-1}\) to a southwesterly wind (13 m s\(^{-1}\)). This still supports a wave passage; however, the wave axis must be oriented northeast–southwest (Fig. 17).

Results from the dropsonde–GDAS analysis comparison and the convective characteristics reveal an important relationship between the strength/scale (organization) of the wave and the accuracy of the GDAS analysis. Dropsonde observations show that waves 1 and 4 are examples that are generally disorganized with low-amplitude wave troughs (“weak waves”); the GDAS analysis is not robust in these cases. This is consistent with large intense raining areas associated with mesoscale circulations that the GDAS analysis is not expected to resolve and that mask the synoptic-scale features of the wave. Although cyclogenesis from waves 1 and 4 is uncertain, their characteristics may exemplify cases that may be more likely to develop owing to stronger incipient vorticity centers and greater latent heating from persistent convection (i.e., wave 2–Debby). In contrast, dropsonde data from waves 3 and 6 show organized low-level synoptic-scale circulations and high-amplitude midlevel troughs that are robust and easy to track in the GDAS analysis; however, waves 3 and 6 are not observed to develop. A common characteristic for waves 3 and 6 is a general absence of large persistent raining areas, consistent with the likely effects of the SAL in inhibiting widespread rainfall.

6. Conclusions

The NAMMA field campaign in the east Atlantic during 2006 provides an unprecedented opportunity to

<table>
<thead>
<tr>
<th>Wave</th>
<th>Total (3-h) snapshots</th>
<th>Fraction of total pixels that are raining</th>
<th>Fraction of raining pixels (≥5 mm h(^{-1}))</th>
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<tr>
<td>1</td>
<td>45 (29)</td>
<td>0.20 (0.01)</td>
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<tr>
<td>2</td>
<td>61 (57)</td>
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<td>0.10 (0.10)</td>
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<td>85 (109)</td>
<td>0.08 (0.07)</td>
<td>0.09 (0.10)</td>
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<td>4</td>
<td>37 (29)</td>
<td>0.20 (0.04)</td>
<td>0.09 (0.18)</td>
</tr>
<tr>
<td>5</td>
<td>57 (57)</td>
<td>0.11 (0.09)</td>
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<tr>
<td>6</td>
<td>45 (—)</td>
<td>0.07 (—)</td>
<td>0.07 (—)</td>
</tr>
<tr>
<td>7</td>
<td>141 (121)</td>
<td>0.24 (0.24)</td>
<td>0.11 (0.14)</td>
</tr>
</tbody>
</table>
examine actual wind and thermal structures at multiple levels in consecutive easterly waves leaving the coast of Africa: an otherwise data sparse region along the propagation path of AEWs. Whereas other studies have used, individually or in combination, rawinsonde, satellite, and operational analysis and reanalysis to characterize key features of AEWs, this study also uses aircraft data, which is mostly lacking in this region. Dropsonde data reveal that wave-to-wave characteristics are widely varying from low-amplitude (weakly organized) to high-amplitude (well organized) wave troughs and are characteristically rainy or dry, depending on the fractional area occupied by the SAL. The ability of the GDAS analysis to subsequently resolve important wave features depends on those characteristics. Waves that are of high amplitude and have well-defined low-level circulations may contain less rainfall and are well resolved in the GDAS analysis but do not necessarily develop. In contrast, waves of low amplitude with disorganized wind structure may have strong vorticity centers and large persistent raining areas and may be more likely to develop.

The relative unimportance of the low-level northern track disturbances has been confirmed; they are characteristically dry and quickly weaken after exiting the coast. Rather, low-level vorticity maxima in the southern track, which develop at the coast or in the ITCZ, may be more relevant for downstream cyclogenesis. In waves 1 and 4, the original wave troughs outrun the vorticity maxima and are no longer present when cyclogenesis occurs. The question remains: What is the relative role of the synoptic-scale wave trough and vorticity maxima in cyclogenesis? Finally, all of the previous results of this study support two important conclusions: one must evaluate multiple pressure levels when studying easterly waves (now common methodology) as well as separate vorticity maxima in the northern and southern track and distinguish between the wave and vorticity maximum scales (Thorncroft and Hodges 2001; Hopsch et al. 2007; Kerns et al. 2008).

While meridional wind and vorticity are the primary variables used to track waves in this study, other candidates for tracking waves include quasi-conserved fields such as potential vorticity or potential temperature and pressure-level data such as geopotential height.

While the extensive dataset of NAMMA provides unprecedented snapshots of both developing and non-developing easterly waves in the east Atlantic, the major question of why one wave develops and others do not remains unanswered. Key snapshots of convection pre- and post-cyclogenesis from fortuitous overpasses of TRMM and other microwave instruments provide an additional focal point for addressing the relationship between easterly waves and cyclogenesis. The problem is dominated by scale interactions and this point has been further investigated for the seven case studies of NAMMA. Understanding how the synoptic scale drives the subsynoptic scale and vice versa has not yet been elucidated and must be explored in future research.

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References

<table>
<thead>
<tr>
<th>Wave</th>
<th>Fraction of total pixels IR</th>
<th>Mean $T_b (K)$</th>
<th>Mean minimum $T_b (K)$</th>
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<tr>
<td>1</td>
<td>0.13 (0.001)</td>
<td>273 (294)</td>
<td>197 (235)</td>
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<tr>
<td>2</td>
<td>0.09 (0.009)</td>
<td>276 (277)</td>
<td>205 (206)</td>
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<tr>
<td>3</td>
<td>0.03 (0.002)</td>
<td>286 (289)</td>
<td>213 (219)</td>
</tr>
<tr>
<td>4</td>
<td>0.14 (0.002)</td>
<td>270 (288)</td>
<td>193 (212)</td>
</tr>
<tr>
<td>5</td>
<td>0.08 (0.005)</td>
<td>275 (282)</td>
<td>199 (202)</td>
</tr>
<tr>
<td>6</td>
<td>0.04 (—)</td>
<td>282 (—)</td>
<td>205 (—)</td>
</tr>
<tr>
<td>7</td>
<td>0.15 (0.017)</td>
<td>260 (265)</td>
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| 1 | The 700 hPa (925-hPa) cold cloud statistics. | | |
|---|---------------------------------------------|---|---|---|---|
| Wave | Fraction of total pixels IR | Mean $T_b (K)$ | Mean minimum $T_b (K)$ |
| 1 | 0.13 (0.001) | 273 (294) | 197 (235) |
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| 6 | 0.04 (—) | 282 (—) | 205 (—) |
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