Numerical Investigations on the Formation of Tropical Storm Debby during NAMMA-06

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ABSTRACT

Mesoscale model forecasts were carried out beginning at 0000 UTC 19 August for simulating Tropical Disturbance 4, which was named Tropical Storm Debby on 22 August 2006. The Weather Research and Forecasting model, with 25-km grid spacing and an inner nested domain of 5-km grid spacing, was used. The development of a small closed vortex at approximately 0600 UTC 20 August 2006 at 850 hPa was found off the coast of Guinea in agreement with satellite images in the 5-km simulation. Intense convection offshore and over the Guinea Highlands during the morning of 20 August 2006 led to the production of a vortex formation by 1400 UTC at 700 hPa. Sensitivity tests show that the Guinea Highlands play an important role in modulating the impinging westerly flow, in which low-level flow deflections (i.e., northward turning) enhance the cyclonic circulation of the vortex formation. Yet, the moist air can be transported by the northward deflection flow from lower latitudes to support the development of mesoscale convective systems (MCSs). Although the model forecast is not perfect, it demonstrates the predictability of the formation and development of the tropical disturbance associated with the Guinea Highlands.

1. Introduction

Our current understanding of tropical cyclogenesis (TC-genesis in the extreme eastern Atlantic remains an important challenge because of the potential dangers that are posed to downstream communities in the Caribbean, Central America, and the United States. Developing tropical cyclones also pose a danger to commercial shipping, local fishermen, and ferry services in coastal regions of western Africa. Most tropical cyclones form from an intensification of the African easterly waves (AEWs), which propagate westward from western Africa toward the North Atlantic and the Caribbean Sea in the range of 10°–20°N (Hopsc et al. 2007; Avila and Pasch 1992; Pasch and Avila 1994). For instance, during 1999, Tropical Storm Cindy formed off the coast of Senegal (Sall and Sauvageot 2005) with reports of more than 100 fishermen fatalities in Senegal, Gambia, and Mauritania. In general, most convective activity [i.e., ensembles of more or less organized short-lived convective cells and mesoscale convective systems (MCSs) of longer duration] weakens when AEWs reach the Atlantic coast, although some MCSs strengthen, resulting in the tropical cyclogenesis stage. Generally, tropical cyclogenesis does not result directly from the synoptic-scale AEWs, but rather from the merger of embedded mesoscale vortices that amplify over the continent in the AEW structure. Other synoptic-scale features including monsoon flow, northeasterly trade winds, Saharan anticyclonic, dry air, and upper-level troughs can also favor or prohibit such an evolutionary pattern. In addition, elevated regions (e.g., Guinea Highlands) have been identified as an important source of MCS generation (e.g., Berry and Thorncroft 2005).

Sall and Sauvageot (2005) described the cyclogenesis that led to the formation of Tropical Storm Cindy (1999) from a mesoscale convective system (MCS) off the coast of Senegal. The tropical cyclogenesis was coincident with an MCS and elements of the easterly wave, that is, a wide area of precipitable water vapor, strong convergence in the low- and midtropospheric layers, and an easterly vertical shear of the zonal wind. Thus, before moving
away from the coast of Senegal this perturbation was named Tropical Storm Cindy.

The AEW life cycle can be categorized into three different phases: initiation, the growth of the AEW over western Africa that results from baroclinic and barotropic processes (Thornicroft 1995), and coastal development (Berry and Thornicroft 2005). The perturbation to the low-level potential temperature gradient becomes severely distorted near the western African coast in the third phase. The westward-propagating potential vorticity (PV) maxima are generated when a strip of PV interacts with the stationary PV maxima over the Guinea Highlands. The merger of these features was responsible for tropical cyclogenesis and later the formation of Hurricane Alberto in 2000 (e.g., Berry and Thornicroft 2005).

In the case of tropical cyclogenesis, it is also important to examine the role of the Guinea Highlands and how MCS evolution is altered in the transition from continental to maritime environments. Further, a detailed analysis of the coherent structures associated with the MCS transition (i.e., structure and evolution) from continental to maritime is needed. Additionally, the Saharan air layer (SAL; Carlson and Prospero 1972) may influence tropical cyclogenesis through aerosol–cloud microphysics interactions leading to invigorated convection (e.g., Koren et al. 2005; Khain et al. 2005; DeMott et al. 2003; Jenkins et al. 2008). The SAL can also have a detrimental effect through the presence of dry air in the midtroposphere, larger static stability, and wind shear (e.g., Dunion and Velden 2004). Jenkins and Pratt (2008) summarized the possible effects of Saharan dust on the development of Tropical Depressions (TDs) 4 and 8 in the eastern tropical Atlantic during 2006. They concluded that Saharan dust aerosols are directly or indirectly influencing tropical cyclone rainbands through increased latent heat release. Both TD-4 and TD-8 are associated with (SAL) outbreaks.

During August and September of 2006, the National Aeronautics and Space Administration (NASA) investigated the downstream aspects of the international African Monsoon Multidisciplinary Analyses (AMMA) campaign (Redelsperger et al. 2006) which was denoted as NASA-AMMA (NAMMA; Zipser et al. 2009). There were several key objectives of the downstream field campaign: characterizations of tropical cyclogenesis, SAL properties, cloud microphysics, precipitation processes, and satellite validation. Aircraft and enhanced ground measurements in western Africa and Cape Verde were used to achieve these objectives. These measurements over the tropical eastern Atlantic were the first since the Global Atmospheric Research Program (GARP) Atlantic Tropical Experiment (GATE) field campaign, where aircraft have successfully investigated tropical cyclogenesis in close proximity to the African continent (Zipser and Gautier 1978).

In this study we will address the following: 1) the structure and evolution of the mesoscale convective system and its connection to TD-4, 2) the impacts of the mesoscale environment (e.g., African easterly Jet, monsoon flow, dry-air intrusions, and vertical shear associated with the SAL) on the evolution of the MCS–vortex, and 3) the role of the Guinea Highlands in the initiation and/or maintenance of the MCS–vortex near the western African coastline. Model forecasts of the AEW associated with Tropical Storm Debby were used to address the items above with an emphasis placed on aspects of the MCS–vortex before tropical cyclogenesis in association with the AEW–MCS transition from continental to oceanic environments from 19 to 21 August 2006.

There is a discussion on the spatial structure and temporal evolution of the pre–TD-4 environment during 19–21 August in section 2. The numerical model and experiment design are described in section 3. The results of our model simulations are discussed and evaluated against observations in section 4, followed by discussion and conclusions in section 5.

2. MCS progression and TD formation

TD 4 developed near 11.6°N and 21.7°W at 1800 UTC 21 August 2006 (Jenkins et al. 2008), from a series of MCSs that propagated westward across western Africa. Figures 1a–d depict the evolution of the MCS using infrared (IR) cloud-top temperatures from 1200 UTC 19 August to 0600 UTC 20 August 2006. The MCS can be traced back to 18 August when it formed near the Chad–Niger region (figure not shown). The MCS cloud tops continued to cool as the system entered Mali on 19 August where the MCS takes on the appearance of a bow-shaped squall line whose intensity decreases as the system moves westward (Figs. 1a and 1b). The average speed of the cold cloud top (<180 K) was about 16 m s⁻¹. The southern part of the MCS weakens, while the northern part continues to propagate westward, crossing Senegal, and reaching the eastern Atlantic at approximately 1800 UTC 19 August (Fig. 1b). Subsequently, the disturbance moved into a dusty environment with large quantities of lightning being observed (cf. Jenkins et al. 2008).

The southern part of the MCS (Figs. 1c and 1d), which produced only small quantities of deep convection embedded in the southwesterly flows, began to show signs of deep convection upon reaching the coastline. In addition to some aerosol–cloud microphysical interactions (Jenkins et al. 2008), the mountainous terrain might have aided the AEW and MCS in the regeneration process through the production of cyclonic motion (vorticity), which was also examined in this study. By 1200 UTC 20 August the deep
FIG. 1. IR brightness temperatures (K) valid at (a) 1200 UTC 19 Aug, (b) 1800 UTC 19 Aug, (c) 0000 UTC 20 Aug, and (d) 0600 UTC 20 Aug 2006.
FIG. 2. As in Fig. 1, but valid at (a) 1200 UTC 20 Aug, (b) 1800 UTC 20 Aug, (c) 0000 UTC 21 Aug, and (d) 0600 UTC 21 Aug 2006.
convection begins to occur off the coast of Guinea (Fig. 2a). Subsequently, from 1800 UTC 20 August to 0600 UTC 21 August (Figs. 2b–d), the system propagated toward the northwest as the spiral cloud band was enhanced. The spiral banding occurred prior to its classification as TD 4 at 1800 UTC 21 August 2006 (not shown).

TD 4 was eventually renamed Tropical Storm Debby as it continued on its path toward the northwest. At approximately 1200 UTC 22 August, the center passed 100 mi southwest of the Cape Verde Islands, bringing thunderstorms and gusty winds. This storm reached a maximum intensity of 50 kt and became a named tropical storm.
(i.e., Debby) near the Cape Verde Islands by 0000 UTC 23 August 2006 (not shown). There were no aircraft measurements of AEW–MCS from August 18 through August 22, as the storm system transitioned from continental to oceanic environments. However, TS Debby was sampled by the NASA DC-8 on 23 August and by the National Oceanic and Atmospheric Administration (NOAA) Gulfstream-IV (G-IV) on 24–25 August with detailed observation and data collections. The storm continued to move WNW and NW at 15–20 kt over the open waters of the eastern Atlantic, with limited intensification due to a dry and stable air mass surrounding the cyclone, and marginal sea surface temperatures. On 25 August, southerly vertical shear began to increase with an upper-level trough, and by 28 August the system dissipated just ahead of the trough (Franklin and Brown 2008). Although the National Hurricane Center (NHC) had forecasted TD 4 to become a hurricane, it never reached hurricane intensity.

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<th>Cases</th>
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<td>TER</td>
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<td>NOTER</td>
<td>25 and 5 (no terrain)</td>
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**FIG. 4.** The 700-hPa streamlines and zonal winds (shaded, m s$^{-1}$) for the CTRL experiment valid at (a) 1200 UTC 19 Aug, (b) 0000 UTC 20 Aug, (c) 1200 UTC 20 Aug, and (d) 0000 UTC 21 Aug 2006.
3. Model description and experiments design

Numerical simulations were performed by using the Weather Research and Forecasting (WRF) model, version 2.2.1 (Michalakes et al. 2005). Two nested domains, as shown in Fig. 3, were constructed with grid spacings of 25- and 5-km horizontal resolution, respectively. The vertical limit of the model was located at the 50-hPa level, and time steps of 60 and 10 s were used in these nested grid simulations, respectively. The 25-km resolution domain covered part of western Africa and the eastern Atlantic (2°–24°N and 31°W–4°E) and the inner domain covers the Guinea Highlands and the extreme eastern portion of the Atlantic (6°–18°N, 23°–6°W). All simulations used the same initial and lateral boundary conditions, which are generated from the National Center for Atmospheric Research (NCEP) Global Forecasting System (GFS) 180-h forecasts with boundary conditions, updated every 3 h at 1° × 1° resolution. The forecast for domain 1 was initialized at 0000 UTC 19 August 2006, and domain 2 was initialized at 0600 UTC, which included enough model spinup time to adapt to the strong diurnal cycle. Both forecasts were integrated until 1200 UTC 22 August 2006.

The planetary boundary layer (PBL) scheme used for this study was the Yonsei University PBL scheme (Hong and Dudhia 2003). The atmospheric radiation scheme accounts for longwave [i.e., the Rapid Radiative Transfer Model (RRTM); Mlawer et al. (1997)] and shortwave (Dudhia 1989) transfers and interactions with the atmosphere, clouds, and the surface. Precipitation was produced from both grid-scale condensation and convection in the 25-km domain. Grid-scale precipitation was determined from an explicit moisture scheme (Thompson et al. 2004) that includes graupel. The atmospheric model was coupled with the Noah land surface model, which includes four soil layers, and it solves the water and energy balance equations at the land surface. The Grell scheme was adopted for cumulus parameterization for...
the 25-km domain, while no cumulus parameterization was used in the inner domain.

The three numerical experiments [the control run (CTRL), and the sensitivity experiments with reduced and zero terrain heights (TER, NOTER)] presented in this study used two-way feedbacks between the large and nested domains (Table 1). We have conducted several sensitivity experiments regarding one- and two-way nested runs (figure not shown) and the best results were found with the two-way feedback option turned on (i.e., the CTRL experiment). In the nested domain of the CTRL experiment, the model used 30-s resolution (≈1 km) terrain data and can account for the possible local/regional circulations due to topography effects. As shown in Fig. 3a, the highest elevations over Guinea Highlands are more than 1300 m. The nested domain of TER used 30-min-resolution (≈54 km) data, and the highest elevation was approximately 300 m over the Guinea Highlands (Fig. 3b). In the nested domain of NOTER, all topography was removed to further illustrate the role of the Guinea Highlands.

4. Results

a. The structure and evaluation of MCSs and TD-4 formation

Figure 4 shows the 700-hPa streamlines and zonal wind at a 12-h interval from CTRL at 25-km grid spacing. An AEW identified at 0000 UTC 19 August as an inverted trough was located near the Gulf of Guinea coastline around 10°N but extended farther westward at slightly higher latitudes (not shown). The African easterly jet (AEJ) during the experiment was located between 14° and 16°N. Twelve hours later, a well-defined AEW was simulated over Burkina Faso (centered on 10°N, 5°W) in the CTRL experiment (Fig. 4a). The inverted trough propagated westward throughout the day and was located off the coastline at 5°–10°N, 15°–20°W but was located farther westward at higher latitudes at 0000 UTC 20 August (Fig. 4b). On 1200 UTC 20 August, as shown in Fig. 4c, a cyclonic circulation associated with the AEW was forming off the coastline. By 1800 UTC 20 August, a closed circulation was simulated. The two-way feedback
with the nested domain provided better scale structure and evolution when compared to the no-feedback coarse domain (not shown). Moreover, the southern part of the AEW was aligned with the northern part of the AEW. However, there is an eastward tilt with latitude at 0000 UTC 21 August (Fig. 4d).

To further investigate the impacts of the mesoscale environment, such as dry-air intrusions, the model-simulated relative humidity and wind vectors at 900 hPa (shown in Fig. 5) were examined. At 1200 UTC 20 August, as shown in Fig. 5a, strong easterly wind existed near 13°N right off the coastline, which was associated with the AEJ (cf. Fig. 4c), with drier air (i.e., RH < 80%) reaching 10°N (Fig. 5a). Another noteworthy feature was a westerly flow that formed around 8°N and impinged on the Guinea Highlands; subsequently, by 1800 UTC 20 August (Fig. 5b), the split flow occurred near the Guinea Highlands. The northern component of the impinging flow formed a cyclonic-like circulation near 10°N, 14°W. A closed cyclonic circulation was simulated near the coast of Guinea Bissau at 0000 UTC 21 August (Fig. 5c), which was consistent with the IR imagery (cf. Fig. 2c). The RH distribution implied that the dry air was associated with the outbreak of Saharan dust, and near the newly formed cyclonic circulation. Six hours later, as shown in Fig. 5d, a well-organized cyclonic circulation was off the coast of Guinea Bissau. This cyclonic circulation brought in moist air (e.g., RH > 80%) from the low latitudes, which supported the cyclogenesis process. The simulation results suggest that a tropical cyclogenesis process occurred within 24 h. The southwesterly flow deflection by the Guinea Highlands mainly occurred at the lower levels.
(i.e., surface \( \sim 1 \) km). Consequently, the AEW trough enhanced the cyclonic circulation development associated with positive potential vorticity (not shown). Overall, three important factors including the AEW, southwestward flow deflection due to the Guinea Highlands, and a moist flow of air being transported from low latitude into continental regions were conducive to the genesis of TD 4. Further discussion of these factors using the nested domain results will follow.

The nested domain (i.e., 5-km grid spacing) started at 0600 UTC 19 August. The simulated 850-hPa streamlines and radar reflectivity (dBZ) of the CTRL experiment from 1200 UTC 20 August to 0600 UTC 21 August are shown in Fig. 6. During the first 24 h of the simulations, the model was able to recapture a westward-propagating easterly wave centered near 3°E with an anticyclonic circulation off the coast of Guinea (9°N, 15°W). By 1800 UTC 19 August (not shown), the CTRL simulation also showed a squall-line feature propagating and exiting the coastline that was similar to the observations at 1800 UTC (cf. Fig. 1b). During the following 12 h from 0600 to 1200 UTC 20 August, the model results demonstrated that the small-scale closed circulation was embedded within an overall cyclonic circulation. A developing cyclonic circulation was found off the coast of Guinea Bissau at 1200 UTC 20 August (Fig. 6a). From 1800 UTC 20 August to 0600 UTC 21 August (Figs. 6b–d), the CTRL experiment demonstrated a broad area of cyclonic circulation that was oriented east to west from 15° to 18°W. The circulation remained close to the coastline throughout the day of 20 August before moving westward on 21 August. By 0000 UTC 21 August, the CTRL experiment clearly demonstrated the formation of the cyclonic circulation around the coastal area (Fig. 6c), and was able to recapture the westward propagation in the following 12 h (not shown). Subsequently, by 0600 UTC 21 August (Fig. 6d), the model result showed well-structured convective activity around the circulation center near 16°W. Six hours later, the circulation center had moved to 18°W and stretched northward. At 1800 UTC 21 August, a well-developed closed circulation was simulated. The center was located near 11°N, 20.5°W and had a sea level pressure of 1011 hPa (not shown). The location of the circulation was eastward and slightly southward of the NHC location of 11.6°N, 21.7°W and 1007 hPa.

In summary, simulated squall-line features produced strong surface wind and intense convection. Subsequently, a cyclonic circulation pattern developed between 1200 UTC 20 August and 1800 UTC 21 August in the CTRL experiment. The model-simulated cyclonic circulation evolution was in good agreement with observed cold infrared cloud-top temperatures (cf. Fig. 2d) located off the coast of Guinea Bissau by 1200 UTC, along with enhanced lightning (e.g., Jenkins and Pratt 2008).

b. The impacts of the mesoscale environment

The mesoscale environment discussed in this paper includes the AEJ, vertical wind shear, and dry-air intrusions from the SAL, all of which contribute to tropical cyclogenesis processes. The time evolution of the 700-hPa zonal wind flow in the CTRL experiment is shown in Fig. 7 at 6-h intervals. The AEJ was located at approximately 15°N from 0600 UTC 19 August to 0000 UTC 20 August (not shown). As shown in Fig. 7a, the AEJ moved southward to about 14°N and there was an intensification of the AEJ between 0000 and 1200 UTC 20 August. The northern component of the developing cyclonic center was located near the southern coastline of Guinea (Fig. 7a), which is in agreement with cold infrared cloud-top temperatures (Fig. 2a). In the following 6 h, the cyclonic circulation was enhanced (Fig. 7b), and was then located downstream of the northern trough axis throughout 20 August and there was a westward tilt. By 0000 UTC 21 August, the disturbance was in phase with the northern trough axis (Fig. 7c).
Subsequently, the cyclonic center moved slightly to the northwest as the northern trough began deepening from 0600 to 1800 UTC 21 August (Fig. 7d). The well-developed vortex center was then located at approximately 12°N, 21°W to the east of the northern trough axis at 0000 UTC 22 August (not shown). The vortex position at 700 hPa was located just east of the NHC estimated position at 0000 UTC 22 August 2006 (i.e., 12.0°N, 22.7°W).

The line of convection and associated diabatic heating over central Guinea at 0000 UTC 20 August led to the production of potential vorticity (PV) on the 315-K potential temperature surface (Fig. 8a). From the morning hours to 1200 UTC 20 August (Fig. 8b), the production of PV continued over the Guinea Highlands. Subsequently, it propagated westward to the coast. There was concentrated convection between 0600 and 1400 UTC offshore and over Guinea at approximately 12°N, 17°W, which led to vortex development at 700 hPa (Fig. 7d). At approximately 0300–0600 UTC 21 August, the vortex was aligned with the AEW and PV was oriented in a north–south direction just offshore (not shown).

The simulated vertical wind shear (500–900 hPa) from 2100 UTC 20 August to 0600 UTC 21 August is shown in Fig. 9. The wind shear was calculated using the simple difference between two levels (e.g., DeMaria et al. 2005). Since the general wind profile over western Africa is characterized by a maximum associated with the AEJ near 700 hPa, and the tropical easterly jet (TEJ) in the upper troposphere is generally not a dominate factor in TC genesis, the wind shear used in this study was based...
on 500 and 900 hPa. The enhanced cyclonic circulation was located near the coast of Guinea and was well defined in the lower troposphere (i.e., 900 hPa) at 2100 UTC 20 August (Fig. 9a). The wind shear was strong in general except for a small region of low wind shear near the cyclonic center at 15°W and 10°N. Additionally, the simulation captured airflow deflection due to the blocking of the Guinea Highlands. Based on the Froude number analysis \( F = U/Nh \); \( U \), upstream wind component \( \approx 10 \text{ m s}^{-1} \); \( N \), upstream Brunt–Väisälä frequency \( = [g/\theta_y(\partial \theta_y/\partial z)]^{1/2} \approx 0.011 \text{ s}^{-1} \); and \( h \), averaged mountain height \( = 1000 \text{ m} \), the Froude number in this case \( F \) is 0.90 ± 0.2. This range of \( F \) indicates that the flow passed over the Guinea Highlands while experiencing orographic modulation (cf. Fig. 4 in Berry and Thorncroft (2005)). By 0600 UTC 21 August, as shown in Fig. 9d, a well-developed cyclonic center was near the same location as 6 h before with an increase in wind speed. The small amount of vertical wind shear provided a favorable environment for the development of a tropical depression.

The developing vortex center from 1200 UTC 20 August to 1200 UTC 21 August revealed long-lived convection, which is conducive to TC-genesis. Therefore, we examined vertical cross sections AA′, BB′, and CC′ (shown in Fig. 3a) of the developing vortex at 1200 UTC 20 August and at 0000 and 1200 UTC 21 August in the CTRL experiment. These three cross sections provided an example of the simulated intensity of deep convection with simulated radar reflectivities greater than 40 dBZ and midlevel

![Diagram of vertical cross sections](image-url)
vertical velocities of 4–12 m s\(^{-1}\). At 1200 UTC 20 August, the small vortex was embedded within the larger trough (cf. Fig. 6a). The cross section was constructed between 8° and 11°N at 14.3°W (AA’, as shown in Fig. 3a), which showed an extension of the AEJ reaching to approximately 10.5°N with low-level westerly winds in the southern part of the vortex (Fig. 10a). A strong band of southwesterly winds (>8 m s\(^{-1}\)), accompanied by strong vertical velocities (>4 m s\(^{-1}\)), were the main features at that time (Figs. 10b and 10c). The relative humidity was greater than 80% overall. High simulated reflectivity (>45 dBZ) is found between 9° and 9.5°N (Fig. 10d). Further analyzing the vertical hydrometeors distributions, the deep convection was associated with the vertical transport of water vapor (not shown). The high values of the reflectivity were associated with rainwater \((q_r)\) mixing ratios in excess of 4 g kg\(^{-1}\) from the ground to near 500 hPa (Fig. 10d). Moreover, the model simulated mid- to upper-tropospheric strong vertical velocities that are associated with latent heat release in excess of 20°C h\(^{-1}\) (figure not shown) from snow, graupel, and ice formation. The simulated high values of the snow \((q_s > 4 \text{ g kg}^{-1})\) and graupel \((q_g > 2 \text{ g kg}^{-1})\) mixing ratios, as shown in Fig. 10d, suggest the importance of these latent heat processes for TC-genesis.

At 0000 UTC 21 August, the vortex began to move northward but was still located near the northern Guinea coastline (Fig. 6c). The cross section was constructed between 9.5°N and 12°N at 15.3°W (i.e., BB’ as shown in Fig. 3). While evidence of the AEJ was still evident (Fig. 11a), strong southerly winds (>15 m s\(^{-1}\)), as shown in Fig. 11b, began pushing bands of convection northward. Several areas of deep convection with high upper-tropospheric vertical velocities (>8 m s\(^{-1}\)) were shown in Fig. 11c. The simulated radar reflectivity of 40 dBZ was found at approximately 350 hPa between 10° to 11°N (Fig. 11d). The water vapor distribution also showed a northward transport of moisture (not shown). The vertical

![Figure 11](image-url)
distribution of $q_{v}$ and $q_{s}$ also demonstrates the northward movement associated with this vortex center (Fig. 11d). Once again, the strong updrafts ($>8$ m s$^{-1}$) were associated with latent heat release from graupel, ice, and snow. A noteworthy feature was that a thin layer of lower relative humidity ($<70\%$) has pushed toward $11^\circ$N, which suggests the possible intrusion of dry air from the north (Figs. 11b and 11c).

At 1200 UTC 21 August, the vortex moved away from land with the simulated maximum reflectivity values located to the northwest and south of the vortex (not shown). A cross section was constructed between $8^\circ$ and $12^\circ$N at $18.35^\circ$W (i.e., CC' in Fig. 3). The center of the vortex was near $11^\circ$N, with weak easterly winds to the north in the lower troposphere ($1-3$ km) and westerly winds to the south (Fig. 12a). Southerly winds were found in the middle and upper troposphere with northerly winds from the surface to approximately $2.5$ km ($\sim 800$ hPa) (Fig. 12b). The strongest updrafts ($>8$ m s$^{-1}$), as shown in Fig. 12c, were found in the middle to upper troposphere, while vertical towers of strong radar reflectivity were also concentrated from the middle to upper troposphere (Fig. 12d). A noteworthy feature is that the water vapor mixing ratio ($>18$ g kg$^{-1}$) was reduced vertically (not shown) as compared to 12 h earlier. There were several convective cells associated with high rainwater mixing ratios, extending up to 500 hPa (Fig. 12d). The relative humidity greater than 95\% was associated with the strong updrafts, as shown in Fig. 12c. Overall, the strong updraft was associated with greater values of graupel, ice, and snow.

c. The role of the Guinea Highlands

Two sensitivity experiments (i.e., TER and NOTER) were conducted to understand the role of the Guinea Highlands in association with the development of MCSs along the coastline. As described in section 2, the topography height in the TER experiment was lower than
that in the CTRL experiment. Figures 13a and 13b show the evolution of the 900-hPa winds and vertical wind shear (500–900 hPa) from 2100 UTC 20 August to 0300 UTC 21 August. During this 6-h period, a weak cyclonic circulation developed near the coast of Guinea, which was due in part to the fact that the AEW had further progressed and was tilted from the SW poleward to the NE (figure not shown). However, as shown in Figs. 13a and 13b, the west to southwesterly flow south of 10°N moved across the Guinea Highlands and penetrated farther inland (cf. Figs. 9a and 9c). In general, weak vertical wind shear zone existed along the coastal region. The NOTER experiment, as shown in Figs. 13c and 13d, from 2100 UTC 20 August to 0300 UTC 21 August showed much stronger easterly winds along the coastline of West Africa. Yet, southwesterly flows dominated the original Guinea Highlands region. The overall wind speed at 900 hPa increased due to the lack of the topography. By 0300 UTC 21 August (Fig. 13d), a small cyclonic circulation occurred near 10°N, 16°W. However, the strong vertical shear zone expanded and propagated westward. When compared to Figs. 9a–d (i.e., CTRL), the TER and NOTER experiments demonstrated different circulation patterns.

In addition to the blocking–deflection effect, the Guinea Highlands also modified the moisture distribution, which could be thus displaced from the already cyclonic vorticity-rich convective region, unfavorable to the tropical cyclogenesis. This can be further explained using the differences in the water vapor \( (q_v) \) mixing ratio at 900 hPa among the CTRL, TER, and NOTER experiments from 1200 to 1800 UTC 20 August. The difference in \( q_v \) between...
CTRL and TER at 1200 and 1800 UTC 20 August demonstrates higher $q_v$ produced near the coastal region from the CTRL experiment (Figs. 14a and 14c). The $q_v$ difference between CTRL and NOTER indicates that higher $q_v$ from NOTER existed along the coastal region (approximately 10°N, 15°W) at 1200 UTC 20 August (Fig. 14b). Nevertheless, by 1800 UTC 20 August (Fig. 14d), the NOTER experiment produced a significantly drier region along 15°W from 10° to 12°N. It appears that the high moist conditions were pushed westward in association with the stronger easterly winds in the NOTER experiment (cf. Fig. 13c). Further, neither sensitivity experiment was able to simulate the pre-TD 4 coastal vortex.

Figure 15 shows the model-derived forward parcel trajectory analysis of the CTRL, TER, and NOTER experiments. The air parcel was selected near 9.5°N, 14°W at the 900-hPa level in the southwesterly flow regime, which was considered to be an upstream air parcel to the Guinea Highlands. The trajectory calculation was based on the wind field (i.e., $u$, $v$, and $w$) selected at a specific location. The velocity data were linearly interpolated in time to the trajectory time steps (i.e., 600 s). In this study, the trajectories were calculated from 1200 UTC 20 August to 1200 UTC 22 August. The forward parcel trajectories from these three experiments showed more deflection of the southwesterly flow due to the blocking effects from the Guinea Highlands (Figs. 15a and 15b), which was conducive to the cyclonic circulation formation in association with the AEW near the coast of western Africa. The trajectory from the NOTER experiment shows the air parcel was affected by the easterly flows after it arrived at 12°N. Overall, the reduced (TER) and no topography (NOTER) runs changed the low-level flow patterns. These two experiments demonstrate the importance of topographic modulation effects from the Guinea Highlands on the vortex development leading to tropical cyclogenesis along the western African coastline.
5. Discussion and conclusions

The forecasting of developing tropical cyclones over the extreme eastern Atlantic remains challenging because of limited observations, the Guinea Highlands, and the interaction between mesoscale convective systems and larger-scale AEWs (2500–3000-km wavelength). Here, we have demonstrated the simulation of the disturbance associated with TD 4 (i.e., TS Debby) using 25- and 5-km grid spacings in the WRF model. This simulation, although not perfect, shows the predictability for the formation of the disturbance in a 78-h forecast. The simulation also confirmed the importance of using two-way feedback, which improves the forecast of the coarse domain as well as the inner domain via time-dependent lateral boundary conditions.

The CTRL simulation first forecasted a squall line that propagated ahead of the developing AEW and a small disturbance within a larger AEW that began to form during the early morning hours of 20 August. By 1200 UTC, a well-defined vortex off the coast of southern Guinea had formed, in good agreement with IR and visible images. Simulated intense convection over Guinea and just offshore during the morning of 20 August was what led to the vortex development off the coast of Guinea. These results suggest the importance of the Guinea Highlands in agreement with the findings from Berry and Thorncroft (2005) for tropical cyclogenesis. The passage of the squall line and its associated stratiform rain on 19 August and the redevelopment of convection over the Guinea Highlands (Figs. 1 and 2) were indirectly (directly) responsible for the production of potential vorticity in the low (mid-) troposphere ahead of the AEW. In this case study, we have shown that the blocking (i.e., flow deflections) effects from the highlands make an important contribution to the vortex development off the coast of Guinea. The Froude number was about 0.90 ± 0.2 in this case. Simulations at later forecast times (i.e., after 0600 UTC 20 August) showed the development of tropical storm strength winds by 22 August (not shown) and demonstrated the WRF’s ability to forecast tropical cyclone development for this case.

However, the question of Saharan dust and its interaction with a developing vortex remains unanswered. Koren et al. (2005) suggest that the suppression of warm rain delays precipitation, with smaller droplets freezing at higher altitudes, increasing the latent heat and leading to stronger updrafts. Jenkins et al. (2008) using satellite observations for the pre–TS Debby environment and limited aircraft measurements for TD 8 suggested that aerosols invigorated the convection. Elevated aerosol optical thickness (AOT) values that extended south toward 10°N are shown in Jenkins et al. (2008) (cf. Fig. 3a) for 20 August. Visible satellite images at 1200 UTC 20 August showed that dust located to the north of the developing convection (not shown). The 1200 UTC cross section between Dakar, Senegal, and Sal, Cape Verde, shows 40%–50% relative humidity between 850 and 700 hPa (Fig. 16). This result is consistent with the simulated relative humidity, which showed an east-west-elongated area of low relative humidity (<40%) at

FIG. 15. The model-derived parcel trajectory analysis from 1200 UTC 20 Aug to 1200 UTC 21 Aug for the (a) CTRL, (b) TER, and (c) NOTER experiments.
approximately 20°N (cf. Fig. 5). This pattern of low RH was located poleward of the observed dust plume and was more elevated than the observed RH between the two cities. Throughout the simulation, dry air could be found to the north and eventually to the west of the vortex (e.g., Figs. 5 and 10–12).

Even with the success of the WRF simulations, uncertainty remains because of aerosols and their potential interactions with the pre–TS Debby disturbance. As suggested by Jenkins and Pratt (2008), the large numbers of cloud-to-ground lightning strikes, high ice content, and latent heat release associated with pre–TS Debby aerosol–microphysics interactions are significant. Additional analysis of NAMMA observations, future field campaigns, and modeling studies with aerosol–cloud microphysics interactions (Van den Heever et al. 2006; Zhang et al. 2007) can help our understanding of how aerosols and their interactions with microphysics processes may be properly parameterized in mesoscale models.

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Fig. 16. Vertical cross section of RH (%) and wind profiles (m s⁻¹) between Sal, Cape Verde, and Dakar, Senegal (~450 km), at 1200 UTC 20 Aug.


