Simulation and Interpretation of the Genesis of Tropical Storm Gert (2005) as Part of the NASA Tropical Cloud Systems and Processes Experiment

SCOTT A. BRAUN
Laboratory for Atmospheres, NASA Goddard Space Flight Center, Greenbelt, Maryland

MICHAEL T. MONTGOMERY
Naval Postgraduate School, Monterey, California, and NOAA/AOML Hurricane Research Division, Miami, Florida

KEVIN J. MALLEN
Department of Atmospheric Sciences, Colorado State University, Fort Collins, Colorado

PAUL D. REASOR
Department of Meteorology, The Florida State University, Tallahassee, Florida

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ABSTRACT

Several hypotheses have been put forward for the mechanisms of generation of surface circulation associated with tropical cyclones. This paper examines high-resolution simulations of Tropical Storm Gert (2005), which formed in the Gulf of Mexico during NASA’s Tropical Cloud Systems and Processes Experiment, to investigate the development of low-level circulation and its relationship to the precipitation evolution. Two simulations are examined: one that better matches available observations but underpredicts the storm’s minimum sea level pressure and a second one that somewhat overintensifies the storm but provides a set of simulations that encapsulates the overall genesis and development characteristics of the observed storm. The roles of convective and stratiform precipitation processes within the mesoscale precipitation systems that formed Gert are discussed. During 21–25 July, two episodes of convective system development occurred. In each, precipitation system evolution was characterized by intense and deep convective upward motions followed by increasing stratiform-type vertical motions (upper-level ascent, low-level descent). Potential vorticity (PV) in convective regions was strongest at low levels while stratiform-region PV was strongest at midlevels, suggesting that convective processes acted to spin up lower levels prior to the spinup of middle levels by stratiform processes. Intense vortical hot towers (VHTs) were prominent features of the low-level cyclonic vorticity field. The most prominent PV anomalies persisted more than 6 h and were often associated with localized minima in the sea level pressure field. A gradual aggregation of the cyclonic PV occurred as existing VHTs near the center continually merged with new VHTs, gradually increasing the mean vorticity near the center. Nearly concurrently with this VHT-induced development, stratiform precipitation processes strongly enhanced the mean inflow and convergence at middle levels, rapidly increasing the midlevel vorticity. However, the stratiform vertical motion profile is such that while it increases midlevel vorticity, it decreases vorticity near the surface as a result of low-level divergence. Consequently, the results suggest that while stratiform precipitation regions may significantly increase cyclonic circulation at midlevels, convective vortex enhancement at low to midlevels is likely necessary for genesis.

1. Introduction

Large-scale influences on tropical cyclogenesis have been studied for many years. There is general agreement that tropical cyclones form in the tropics or subtropics over sufficiently warm (>26°C) water possessing sufficiently great depth, far enough from the equator that background rotation is sufficient, in regions of high relative humidity, and when vertical wind shear over the depth of the troposphere is relatively small (Gray 1975, 1979). In addition, tropical cyclones form within regions of pre-existing cyclonic relative vorticity in the lower
troposphere—for example, easterly waves, a monsoon trough, or the active part of the Madden–Julian oscillation (Roundy and Frank 2004).

Tropical storms are generally spawned from meso-scale convective system (MCS) precursors within the pre-existing region of cyclonic vorticity noted above (Velasco and Fritsch 1987; Gray 1998). Midlevel convergence into large stratiform precipitation regions within MCSs, along with tilting of horizontal vorticity into the vertical, provides a source of concentrated midlevel vorticity (Gamache and Houze 1982; Verlinde and Cotton 1990; Brandes and Ziegler 1993; Chen and Frank 1993; Bister and Emanuel 1997; Chong and Bousquet 1999; Yu et al. 1999) that often becomes the precursor to surface development. Although MCSs occur frequently over the tropical oceans, only a few develop into tropical cyclones and the mechanisms that inhibit or favor development are still poorly understood. Over the last decade and a half, the focus on the genesis problem has been the search for a mechanism responsible for the development of low-level vorticity below the MCS of sufficient intensity to initiate the wind-induced surface heat exchange (WISHE) process of Emanuel (1986, 1987). Several studies have proposed mechanisms for the generation of sufficient surface cyclonic vorticity by some form of vorticity transport or projection downward from the midlevels. These are the so-called “top-down” theories of Bister and Emanuel (1997), Ritchie and Holland (1997), and Simpson et al. (1997). Bister and Emanuel (1997) proposed that a mesoscale region of light rainfall, or stratiform rain, would act to humidify the low-level air, thereby gradually lowering the level of peak cooling—and hence potential vorticity (PV) production—to the surface. The key element in this hypothesis is the requirement of a stratiform precipitating region that cools and moistens the lower troposphere and descent of the cyclonic vortex to near the surface to the point at which the effects of cold downdrafts no longer inhibit development of cyclonic winds at the surface. Ritchie and Holland (1997) and Simpson et al. (1997) proposed a vortex merger theory in which successive mergers of midlevel mesoscale vortices (generally thought to be associated with the stratiform regions of MCSs) intensified the midlevel vortex. A consequence of the midlevel merger process is an increase in the horizontal and vertical scale of the vortex. They proposed that genesis would begin when the vertical scale had increased sufficiently to reach the surface.

Hendricks et al. (2004) and Montgomery et al. (2006) have proposed an entirely different “bottom-up” deep-convection route to cyclogenesis that blends moist thermodynamic and dynamic processes and operates between the development of a weak cyclonic circulation near the sea surface and the ignition of the WISHE mechanism. In their high-resolution numerical simulations, Montgomery et al. (2006) found that deep convective towers possessing intense cyclonic vorticity in their cores are the dominant coherent structures of a predepression disturbance. These vortical hot towers (VHTs) sustain themselves by consuming available potential energy in their local environment and by merging with neighboring towers. The population of VHTs statistically mimics a quasi-steady heating rate in the core of the mesoscale vortex and generates a system-scale transverse circulation with low-level inflow and upper-level outflow. The low-level inflow concentrates the pre-existing and VHT-generated absolute cyclonic vorticity to a sufficient amplitude to start the hurricane heat engine.

Tory et al. (2006a,b), using output from the Tropical Cyclone Limited Area Prediction System (TC-LAPS, with 0.15° horizontal resolution), determined that the primary vortex enhancement mechanism in the model was convergence/stretching of absolute vorticity in deep convective updrafts. Secondary vortex enhancement mechanisms were associated with vortex upscale cascade, or mergers of multiple convective vortices into a single larger vortex, and system-scale intensification via enhancement of the secondary circulation by convective heating. They argued that while stratiform precipitation regions may significantly increase cyclonic circulation at midlevels, convective vortex enhancement at low to midlevels is likely necessary for genesis. However, stratiform precipitation was largely absent from the TC-LAPS simulations because of the coarse resolution and lack of explicit cloud microphysical processes. Given the use of a convective parameterization, the vertical motions were dominated by large deep convective cores that may have likely biased the divergence profiles toward convective rather than stratiform profiles.

Given these uncertainties regarding the relative roles of convective and stratiform precipitation regions in cyclogenesis, in July 2005 the National Aeronautics and Space Administration (NASA) conducted the Tropical Cloud Systems and Processes (TCSP) field experiment in collaboration with the National Oceanic and Atmospheric Administration’s (NOAA) Hurricane Research Division (HRD) to study tropical cloud systems and tropical cyclone genesis and evolution in the eastern Pacific and western Caribbean (Halverson et al. 2007). A major objective of the TCSP experiment was the improvement of the understanding and prediction of tropical cyclone genesis using remote sensing and in situ data, as well as numerical modeling, particularly as they relate to the three phases of water and the organization of precipitation. On 23–25 July, the NASA ER-2 and NOAA P-3 aircraft flew repeated missions into a tropical wave that
eventually transformed into Tropical Storm Gert before making landfall in Mexico along the western Gulf coast. A future paper by Mallen et al. will present a detailed observational analysis of the formation and evolution of Gert. This study describes a numerical modeling study of the genesis of Gert, with an emphasis on the evolution of its precipitation, kinematic, and thermodynamic structures. Specifically, we seek to elucidate the roles of well-resolved convective and stratiform precipitation processes in the generation of potential vorticity within the storm and the development of surface circulation leading to genesis.

2. Methodology and data description

a. Model setup

This study employs the Advanced Research version of the Weather Research and Forecasting (WRF) modeling system (version 2.2; Skamarock et al. 2005) to conduct simulations of the genesis of Tropical Storm Gert. Four grids nesting down to 2-km horizontal grid spacing (Fig. 1) are employed to adequately represent the convection. The outer grid has a horizontal grid spacing of 54 km and contains 150 × 90 grid points in the x and y directions. The grid is centered at 22.9°N, 91.1°W and uses a Mercator map projection. Two stationary nested meshes are used with the following grid spacings and grid dimensions: 18 km and 226 × 178, and 6 km and 400 × 340. The fourth nest is designed to move with the storm and has a grid spacing of 2 km and dimensions of 400 × 400 grid points. All grids use 31 vertical levels. Physics options include the Yonsei University boundary layer scheme (Noh et al. 2003; Hong et al. 2006), the fifth-generation Pennsylvania State University–National Center for Atmospheric Research (NCAR) Mesoscale Model (MM5) similarity-theory surface-layer scheme (Zhang and Anthes 1982; Skamarock et al. 2005), the Noah land surface scheme (Chen and Dudhia 2001), the Kain–Fritsch cumulus scheme (Kain and Fritsch 1990, 1993; Skamarock et al. 2005) on the 54- and 18-km grids only and calculated every time step, and the WRF single-moment six-class cloud microphysics (Hong et al. 2004) on all grids. Radiative processes are calculated every 5 min on the 54- and 18-km grids and 2 min on the 6- and 2-km grids using the Rapid Radiative Transfer Model (RRTM) longwave (Mlawer et al. 1997) and Dudhia shortwave (Dudhia 1989) schemes.

Initial and boundary conditions are obtained from 6-hourly National Centers for Environmental Prediction (NCEP) Global Forecast System (GFS) analyses with 1° resolution using the WRF preprocessing system software. Experiments were run with multiple initialization times to determine which times provided the best reproduction of the evolution of Gert as verified by aircraft and satellite observations. In this study, results are shown for two simulations. The primary simulation (designated the Control run) is started at 0600 UTC 22 July 2005 and run for 66 h until 0000 UTC 25 July. This simulation verifies well against observations but produces a weaker surface pressure minimum at landfall than is observed. A second simulation (designated the Sim2 run) is started at 1200 UTC 21 July and is discussed in section 5. This simulation produces a stronger vortex and more active convection at early stages (22–23 July) but is less consistent with observations at these times. This second simulation is of interest because it highlights development in the context of a stronger background vortex and leads to a more organized system with minimum surface pressures that are somewhat lower than observed values. While these two simulations clearly do not constitute a large sample of a possible ensemble, they encapsulate the overall genesis and development characteristics observed. Consequently, we believe these two simulations are adequate for answering the primary scientific questions raised in the introduction.

b. TCSP airborne validation data

The NASA TCSP field experiment included research flights with the NASA ER-2 high-altitude aircraft, typically flying at ~20-km altitude, and two NOAA WP-3 Orion aircraft flying near 650 hPa (~3.5 km). For Tropical Storm Gert, five missions were conducted over the life cycle of the storm, from when the disturbance was a tropical wave over the Yucatan Peninsula to shortly after landfall as a tropical storm. Two of the missions involved coordinated flights with the ER-2 and one P-3 aircraft, with the remainder being single aircraft missions. In addition to the flights associated with the TCSP
experiment, an Air Force reconnaissance flight occurred during 0849–1736 UTC 24 July, thus providing greater continuity of measurements during the life cycle of Gert. Table 1 provides a summary of the different flights. See Halverson et al. (2007) for a detailed description of the flights in this case.

Validation efforts focus primarily on the wind information from Doppler radar and dropsondes from the NOAA P-3 aircraft as well as the flight-level winds from the Air Force flight on 24 July. The processing of the radar data volumes is described in Reasor et al. (2009). Several volumes are processed during the period 0326–0600 UTC 24 July and are combined into a larger composite of reflectivity and wind fields assuming a reference time of 0400 UTC and a storm motion of 7.6 m s\(^{-1}\) to the northwest as determined from National Hurricane Center (NHC) best-track information. A Doppler volume near 2100 UTC 24 July was also processed, capturing the developing system just prior to landfall. The P-3 derived radar reflectivities had a clear low bias when compared to reflectivities from the ER-2 Doppler (EDOP) radar, with peak EDOP values being 10 to as high as 20 dB\(Z\) higher than the P-3. The lower reflectivities from the P-3 are likely the result of strong attenuation (Jiang et al. 2006). Therefore, P-3 reflectivities should be viewed only qualitatively here. Dropsonde information is overlaid on the Doppler analyses and is shifted in space assuming the appropriate storm motion and reference times.

3. Simulation results and validation

a. Storm evolution

This section focuses on the evolution of the simulated development of Tropical Storm Gert through a description of the simulated low-level wind and precipitation fields. Figure 2 shows the simulated radar reflectivity\(^1\) and winds barbs at 500 m for selected times. Six hours into the simulation (Fig. 2a), intense deep convection developed along and offshore of the Belize coast, qualitatively similar to Tropical Rainfall Measuring Mission (TRMM)-observed convection at this time (Fig. 3a) except that the west–east-oriented band was observed to be much farther northward. A low-level cyclonic circulation was located along the Honduran coastline at the end of the hook-shaped convective system. At 1800 UTC 22 July (Fig. 2b), the center of circulation was along the Belize coast and convection was beginning to diminish as the system moved inland. As the circulation moved into the southern Bay of Campeche by 1200 UTC 23 July (Fig. 2c), there were scattered areas of convection over the Gulf, much of it fairly shallow (below ~5 km). By 1800 UTC 23 July (Fig. 2d), the circulation continued to drift northwestward as some convection developed north of the center. NHC best-track data designate the disturbance Tropical Depression 7 at this time (it became Tropical Storm Gert by 0600 UTC 24 July), and its position (marked by the X in Fig. 2d) is in good agreement with the simulated wind field. Westerly flow just south of the center was very weak; by 0000 UTC 24 July (Fig. 2e) it was replaced by weak easterly flow such that the closed circulation was absent when viewed in an earth-relative reference frame. In a frame of reference moving with the wave disturbance, however, a closed circulation exists at this time in the lower troposphere (1000–500 mb; not shown). The presence of a closed circulation in the wave frame is thought to be a critical ingredient for a successful wave-to-vortex transformation (Dunkerton et al. 2009). This boundary demarcating the closed circulation is an approximate material boundary that acts over most of the depth of the vortex to reduce dry air intrusion and contain moisture lofted by deep convection, such as VHTs (Dunkerton et al. 2009).

Two regions of more intense precipitation were present at 0000 UTC 24 July. The first was just west of the trough axis along a convergence zone where the northerly flow associated with the trough met a low-level barrier jet east of the mountains. The second was an area of organizing deep convection embedded within the strong southeasterly flow on the northeastern side of the trough. Convection began to rapidly expand by 0600 UTC 24 July (Fig. 2f). In the southwestern Bay of Campeche,

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\(^1\) See footnote 2 of Braun et al. (2006) for a description of the reflectivity calculation.
the easterly flow to the east of the trough axis encountered the northwesterly barrier jet, leading to an enhancement of convergence and convection and reforming the closed cyclonic circulation (in an earth-based reference frame) about 0.25° to 0.5° to the north and west, respectively, of the NHC-estimated storm center location. Convection surrounded the center of circulation with weaker flow within the ring of convection and stronger flow without. By 1200 UTC (Fig. 2g), convection intensified around the center of circulation, which elongated in the north–south direction and shifted southward slightly so that it was on the north side of intense convection that had formed on the border of the barrier jet. Over the next 6 h (Fig. 2h), convection on the east side of the circulation dissipated or moved northward while convection on the west side was enhanced along the coastline. The center of circulation eventually made landfall at approximately 2300 UTC 24 July, within a few hours of the observed landfall at or just after 0000 UTC 25 July.

b. Observational validation

In this section, we provide validation of the simulation using data from the NOAA P-3 Orion and Air Force reconnaissance aircraft flights into Gert, as well as data from the NASA TRMM and QuikSCAT satellites. While a wealth of other data was available, here we show only those data that provide critical validation of key features of the storm’s evolution, with an emphasis on winds and precipitation.

The QuikSCAT satellite passed over the Gulf of Mexico three times during the genesis of Gert. Wind retrieval accuracy is impacted by rainfall as a result of scattering and attenuation of the transmitted energy by rain as well as effects of the rain on the surface roughness of the ocean. In rainy areas, the retrieved wind is generally too large by an amount that is proportional to the rain rate (Portabella and Stoffelen 2001). For rain rates above 6 mm h⁻¹, as determined by Special Sensor Microwave Imager (SSM/I) data, Portabella and Stoffelen (2001) suggested that the QuikSCAT wind vector cells contained no useful wind information. While plots of the QuikSCAT data in Figs. 4–5 include all of the available wind data, areas having more than a 50% probability of rain in the QuikSCAT footprint are indicated to highlight areas of possible rain contamination and overestimation of the wind speed. At 0000 UTC 23 July (Fig. 4a), the QuikSCAT wind field was characterized by a broad area of weak northeastern flow. Wind vectors just off the western coast of the Yucatan Peninsula suggest a cyclonic disturbance over the peninsula. The model wind field at this time (Fig. 4b) is similar to the QuikSCAT wind field. At 1200 UTC 23 July (Fig. 5a), the center of circulation is along the southern coast of the Bay of Campeche between 92° and 93°W and a well-defined convergence zone is present where the flow on the western side of the trough encounters offshore-directed flow, which the model suggests is related to topographic blocking. This pattern is qualitatively similar to the corresponding simulated wind field (Fig. 5b) but with wind speeds that are ~2 m s⁻¹ stronger. Although the wind vectors are to some degree contaminated by rain along the western coast, the QuikSCAT data suggest a greater offshore extent of the barrier flow than seen in the simulation. The QuikSCAT winds generally indicate that the evolution of the surface winds within the model is reasonable at early stages.

Figure 6 shows a comparison of the P-3-derived composite reflectivities, Doppler winds, and dropsonde winds, and the simulated reflectivities and winds at 2- and 4-km altitude. The radar and dropsonde data correspond to a reference time of 0400 UTC 24 July and are compared to model fields at 0600 UTC 24 July as a result of a delay by a few hours in the model of the intensification of the convection, as determined from Geostationary Operational Environmental Satellite (GOES) observations. At both 2 and 4 km (Figs. 6a,b), the radar data indicate scattered areas of convection with intervening stratiform precipitation. The flow is generally southeasterly east of ~95°W and north of 19.5°N, turning to northerly or northwesterly west of 95°W. Data are sparse in the southern portion of the system but suggest an elongated trough or perhaps closed circulation rather than a well-defined, smaller-scale center of circulation. The simulation also shows scattered areas of deep convection, with stratiform precipitation primarily limited to more northern areas. The simulated winds also show an elongated cyclonic circulation. Simulated winds in the northern precipitation area are similar to the observed values at 2 km, but up to a few meters per second weaker at 4 km. To the west, simulated winds at 2 km (Fig. 6c) show the correct wind direction but are about 2–3 m s⁻¹ weaker and occur outside a simulated radius of about 1° compared to an observed radius of about 0.5° (here, radius refers to the distance from the approximate center of circulation). At 4 km (Fig. 6d), the simulated winds are from the north to northeast compared to the observed northerly to northwesterly winds, although the magnitudes are reasonably close to observed. This comparison suggests that the simulated structure is at least qualitatively consistent with the observations, but with winds that are somewhat weaker than observed and with a simulated vortex that is not as well established at mid-levels as in the observations.

Figure 7 shows a similar comparison between the model and observations, but for 2130 UTC 24 July. The simulated storm has made landfall by this time (Fig. 7c),
whereas the observed storm center was located just off-shore (Fig. 7a). The P-3 radar indicates the most intense precipitation to the north of the center, with weaker precipitation around the eastern and southern sides. The simulation shows intense precipitation on the eastern side, weaker precipitation to the south, and also precipitation over land to the west, which was not within the radar range in this Doppler volume. Observed winds at 2 km (Fig. 7a) are up to 20 m s$^{-1}$ to the north of the center and 10–17 m s$^{-1}$ to the east, with wind speeds generally increasing with radius out to about 1° from the circulation center. The simulated winds at 2 km (Fig. 7c) are about 10–15 m s$^{-1}$ directly north of the center, up to 20 m s$^{-1}$ to the northeast, and up to about 18 m s$^{-1}$ to
the east, in good general agreement with the Doppler winds. The simulated and observed winds to the south of the center are also in good agreement, with magnitudes around 7–10 m s\(^{-1}\) within 0.5° of the circulation center.

At 6 km (Fig. 7b), the Doppler winds capture mainly the strong southeasterly flow north and east of the center and very light winds near the center. Although not well defined because of relatively sparse wind data, the circulation center is located closer to the coast, to the south of the 2-km level center, indicating a tilt of the vortex to the south. The simulation (Fig. 7d) shows winds of comparable magnitude, with winds between 15 and 20 m s\(^{-1}\) to the north of the center. The center of circulation is also shifted southward at 6 km compared to 2 km, in agreement with the Doppler observations.

The TRMM satellite passed over Gert at ~1430 UTC 24 July during a time when Air Force reconnaissance was flying in the boundary layer within the storm. A subset of the flight-level winds is overlaid on the TRMM rainfall rates in Fig. 3b. Note that rainfall rates are derived...
from the TRMM Precipitation Radar (PR) within its narrow swath (indicated by thin blue lines), whereas elsewhere the rain rates are retrieved from the TRMM Microwave Imager (TMI). The most intense rainfall is found in the eastern portion of the rainband on the southern side of the storm (20.3°N, 95.75°W). This result is true even if the PR data are excluded (not shown), in which case the TMI rainfall rates in this area would be slightly greater than 20 mm h⁻¹. The Air Force flight-level winds show strong westerly flow within the southern rainband and indicate a center of circulation near the northeast edge of the band near the most intense convection. Farther north is a wide area of lighter precipitation with embedded convective cores (TMI rainfall rates >10 mm h⁻¹), while to the east is a rainband with relatively low rainfall rates (PR rainfall rates <20 mm h⁻¹). Simulated rainfall rates, along with 500-m-level wind barbs for a region comparable to that in Fig. 3b, are shown for the same time in Fig. 3c. In many respects, the model compares quite favorably with the observations. The overall rainfall pattern is quite similar to that observed by TRMM, with a prominent rainband with intense convection on the southern side of the storm, a wide area of precipitation to the north,
and a rainband to the east. Furthermore, the simulated winds clearly indicate a center of circulation coinciding with the northern edge of the southern rainband with strong westerlies within the rainband. We can note also some differences: more intense convection scattered throughout the storm (cf. to rainfall rates within the PR swath), particularly in the north–south-oriented band to the east of the center, and weaker winds in the simulation compared to Air Force measurements, suggesting an underdevelopment of the storm circulation at this time.

Fig. 4. (a) QuikSCAT wind speeds (shading) and vectors for 0100 UTC 23 Jul. Areas enclosed within the solid line have a 50% probability of rainfall contamination. (b) Simulated 10-m winds at 0000 UTC 23 Jul.

Fig. 5. As in Fig. 4, but for 1200 UTC 23 Jul.
As a final comparison, Fig. 8 shows the simulated minimum sea level pressure along with the observed best-track value. The observed value may contain considerable uncertainty given the limited sampling of the storm during its evolution. The observations suggest nearly continuous deepening of the storm from 1011 to 1005 hPa beginning at 1800 UTC 23 July, when Gert became a named storm, and ending at 0000 UTC 25 July, when the storm made landfall. The simulation showed little tendency for deepening until about 0600 UTC 24 July, when significant deep convection began, subsequently deepening from 1012 to 1008 hPa. On the basis of the foregoing findings, we conclude that the storm development was delayed and weaker relative to the observations.

4. The relative roles of convective and stratiform processes

In this section, we investigate the evolution of the mean vortex and the role that convective and stratiform
precipitation processes play in this evolution. Simulation results are examined in a reference frame centered on the storm. When possible, the storm center is estimated using an approach similar to that described in Braun (2002) and Braun et al. (2006), but here minimizing the asymmetry of the 850-hPa geopotential height instead of sea level pressure. Because the storm is weak, there is considerable uncertainty in the center position. At early times, generally prior to 0900 UTC 22 July, the geopotential height field provides a poor indicator of the disturbance position, so the approximate center of the developing convective system is used instead. To minimize the impact of this uncertainty in the center location, the results below show time series of profiles of area-averaged quantities, averaged within a radius of 300 km from the storm center. Fields of tangential and radial velocities must be viewed with caution, especially at earlier times (e.g., prior to 1200 UTC 23 July when storm-generated geopotential height anomalies were weaker), because their values are dependent on the derived center locations.
The tangential flow (Fig. 9b) shows that the cyclonic circulation extended vertically from the surface to between 300 and 400 hPa and had peak tangential velocities while other fields simply reflect the evolution of the convective system and its near environment.

Figure 9 shows time series of the area averages of several quantities for the Control simulation beginning at 0600 UTC 22 July and ending at 0000 UTC 25 July. The vertical motion (Fig. 9a) shows an initial burst of ascent prior to 1200 UTC 22 July as precipitation develops on the eastern side of the Yucatan Peninsula (cf. Fig. 2a). Convection weakened during the later part of 22 July and the early part of 23 July (Figs. 2b,c), transitioning to mean downward motion at middle to upper levels around 1200 UTC 23 July. Convection resumed around 1800 UTC 23 July (Fig. 2d) and then increased substantially by 0600 UTC 24 July (cf. Fig. 2f). After 1200 UTC 24 July, decreasing low-level upward motion and increasing upward motion aloft indicated a growing influence of stratiform precipitation processes.

To better delineate the roles of convective and stratiform processes during the simulation, Fig. 10 shows the area-weighted averages of vertical motion for convective, stratiform, and nonprecipitating (at the surface) regions. Specifically, the area-averaged value of any quantity $a$ can be defined as $\overline{a} = \sigma_c \overline{a}_c + \sigma_s \overline{a}_s + \sigma_e \overline{a}_e$, where subscripts $c$, $s$, and $e$ denote convective, stratiform, and environment; $\overline{a}_c$, $\overline{a}_s$, and $\overline{a}_e$ are the average values in the respective regions; and $\sigma$ is the fractional area (e.g., convective area divided by total area) for each region. The area-weighted average values for each region, as in Fig. 10, are then $\sigma_c \overline{a}_c$, $\sigma_s \overline{a}_s$, and $\sigma_e \overline{a}_e$. The separation into convective and stratiform components was accomplished using a method similar to that of Tao et al. (1993). First, all grid points with surface rainfall rates greater than 20 mm h$^{-1}$ were classified as convective. Next, a texture algorithm was used, whereby grid points having rainfall rates twice as large as the average of their nearest 24 neighbors were classified as convective. If a grid point is designated as convective in this way, then its nearest neighbors (within one grid distance) are also designated as convective. To identify convective columns in which significant precipitation was not yet reaching the surface, columns with upward vertical motions $>3$ m s$^{-1}$ or cloud liquid water $>0.5$ g kg$^{-1}$ were also denoted as convective. All remaining grid columns with surface precipitation greater than 0.1 mm h$^{-1}$ were classified as stratiform, while remaining grid columns were classified as environment or nonprecipitating anvil. The fields shown in Fig. 10 depict the averages over each region weighted by the fraction of the total number of grid columns in each classification. The sum of Figs. 10a–c yields the average vertical motion in Fig. 9a.

Figure 10 shows that within the first 6 h of the simulation, the initial precipitation development is stratiform in character because of the large-scale saturated ascent (resulting in turn from the initialization with coarse fields from the NCEP analysis). Within a few hours, small-scale structure emerges in the form of deep convection that tends to dominate the vertical mass flux until about 1200 UTC 22 July. By that time, a stratiform precipitation region forms in association with the deep convection and is associated with weak upward motion peaking at 300 hPa and weak descent below 550 hPa. Strong subsidence occurs in the environment of the initial convection, with peak downward motion at heights between 300 and 200 hPa. Convection weakens early on the 23rd, with negligible mean ascent or with mean subsidence found at mid to upper levels and with convection (cf. Fig. 2c) generally limited to below 500 hPa. Weak, deep convection develops late on 23 July, with strong deep convection beginning around 0300 UTC 24 July. Mean convective ascent peaks around 1200 UTC 24 July and then weakens, although localized regions of strong ascent continue near the storm center through the end of the simulation. Mean stratiform ascent (Fig. 10b) at upper levels and descent below 500 hPa develops around 2100 UTC 23 July, intensifies rapidly after 0600 UTC 24 July, and peaks near 1500 UTC on the 24th. In the nonprecipitating region (Fig. 10c), mean ascent develops in nonprecipitating anvils after 1200 UTC 24 July.

The tangential flow (Fig. 9b) shows that the cyclonic circulation extended vertically from the surface to between 300 and 400 hPa and had peak tangential velocities.

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2 These criteria should not be viewed as a kinematic or microphysical definition of convective versus stratiform, but simply as a means of diagnosing convection at very early stages that might not be identified from the earlier criteria.

3 When the averaging radius is reduced to 150 km, both convective and stratiform upward motions remain strong through the end of the simulation, suggesting that much of the decrease seen in Fig. 9 occurs in the radial band between 150 and 300 km.
near the top of the boundary layer throughout the simulation. Anticyclonic flow occurred above 300 hPa. The vortex was relatively strong at 1200 UTC 22 July as a result of the convection on the east side of the Yucatan Peninsula prior to that time. As the storm moved over the peninsula during the later part of 22 July and early part of 23 July, both the convection and the tangential velocities weakened. The weakening of the midlevel tangential velocities during this time does not necessarily reflect a weakening of the circulation (notice that the average PV did not change much in Fig. 9f). Instead, it appears to result from a slower westward movement of high-PV air at midlevels (~500 hPa) compared to air at low levels (~850 hPa), thereby producing an eastward tilt of the system. Since the strongest winds were ~350 km northeast of the center during this time, this eastward tilt of the storm resulted in stronger winds at midlevels being shifted out of the averaging domain, thus lowering the average tangential wind speed. The resumption of convection, first just after 1200 UTC 23 July and then more intensely after 0000 UTC 24 July, led to a realignment of the lower and midlevel circulations as well as a simultaneous intensification of the lower and midlevel tangential flow. Beginning around 1200 UTC 24 July, when stratiform ascent was strong, rapid intensification of the midlevel flow began.

The mean radial velocities (Fig. 9c) suggest strong boundary layer and midlevel convergence and upper-level divergence through much of the early stages of convection. With the development of convection late on the 23rd and early on the 24th, low-level convergence and upper-level divergence increased dramatically. In addition, a layer of midlevel to upper-level convergence developed in association with convection late on the 23rd and intensified further after 1200 UTC 24 July as a result of the development of stratiform precipitation. These radial velocity (divergence) profiles are similar to those reported by Montgomery et al. (2006) in their examination of a VHT pathway to tropical cyclone genesis within a parent mesoscale convective vortex (MCV).
shallow layer of inflow in the boundary layer and the finding of peak tangential winds at the top of the boundary layer are consistent with Smith et al. (2009), who argued that convergence of absolute angular momentum in the boundary layer is an important mechanism for spinning up the inner core of a storm.

Relative humidities (Fig. 9d) at low to middle levels were near saturation throughout the simulation and a deep layer of near-saturated conditions existed during the initial spinup of precipitation. The trend toward weakening ascent and the transition to mean descent by 1200 UTC 23 July led to a drying out of the layer above 400 hPa during that time, although lower levels remained nearly saturated. As convection redeveloped late on the 23rd, the upper levels quickly moistened, generally with humidities that approached saturation with respect to ice.

Figure 9e shows the evolution of the potential temperature anomaly. The anomaly is defined as a perturbation with respect to the near-storm environment, determined by averaging the potential temperature within the radial band 300–350 km (in the annulus immediately outside of the 300-km radius averaging area). A warm anomaly is found between 800 and 300 hPa prior to 1200 UTC 23 July, thereafter deepening to 200 hPa. The temperature anomaly becomes largest after 0600 UTC 24 July between 300 and 200 hPa as stratiform precipitation becomes better developed and approximately coincides with the onset of surface pressure falls (Fig. 8). Potential temperatures at low levels remain cool throughout the simulation, with the cool layer deepening (from 800 to 600 hPa) after 1200 UTC 24 July. These results suggest that genesis can occur despite the maintenance of the surface cold pool. Similar results were also observed in the idealized simulations of Montgomery et al. (2006).

The potential vorticity (Fig. 9f) shows a midlevel vortex centered near 600–500 hPa, with very weak PV in the upper troposphere. There is little trend in the PV time series except after 1200 UTC 24 July, when PV increases as significant stratiform precipitation develops. However, although the area-averaged PV is relatively static, significant changes in PV do occur as PV is redistributed by convection, as will be shown in more detail below. Prior to that discussion, let us first look at the mean PV within convective and stratiform regions in order to examine key characteristics in these regions. Figure 11 shows the area-averaged PV in convective and stratiform regions (i.e., $PV_c$ and $PV_s$). Weighting by area, as was done with vertical velocity in Fig. 10, was not performed since the mean in Fig. 9f is dominated by the nonprecipitating environment or anvil region. Throughout the simulation, convective and stratiform regions exhibit specific profiles of PV. In the convective areas, PV is maximum at low levels, typically below 800 hPa, but with high PV extending upward to near 400 hPa. Very low or even negative PV is found in the upper troposphere between 300 and 200 hPa. In stratiform regions, PV is generally maximum at middle levels around 500 hPa, with high values extending down to near the surface and very low values in the upper troposphere. Examination of animations of the PV field suggests that some of the higher-PV air at low levels in the stratiform region originated within convective regions. The results in Figs. 10 and 11 indicate that the stratiform precipitation regions primarily enhance PV at midlevels, supporting the work of Tory et al. (2006a,b). Without deep convection, convergence is limited to midlevels whereas divergence occurs at low levels, which does not favor vorticity enhancement near the surface. As in Tory et al. (2006a), deep convection favors vorticity enhancement at low to midlevels, suggesting that genesis cannot begin without deep convection.

The convectively generated PV at low levels has a large impact on the pressure and wind fields. Figure 12
shows plots of smoothed\(^4\) 850-hPa PV and sea level pressure for selected times. At 1200 UTC 23 July (Fig. 12a), even though there was relatively little deep convection at this time (see Fig. 10a), there were some isolated areas of intense convection, the most prominent being near the center of the storm in the sea level pressure field. A local core of very intense PV (labeled P1) was associated with this convection and had just moved into the Bay of Campeche from the Yucatan Peninsula. Pressure perturbations associated with this intense PV feature added to the pressure deficit present on larger scales so that the pressure minimum was collocated with P1. Six hours later (Fig. 12b), P1 had moved west-northwestward and continued to be collocated with the minimum pressure. By 0000 UTC 24 July (Fig. 12c), the PV maximum P1 was still present, a good 12 h after its formation, and continued to be associated with the minimum pressure. A second intense PV feature (labeled P2) formed to the northwest of P1. Over the next 4 h (Fig. 12d), P1 weakened. From this point on, although the original PV maximum weakened, the pressure minimum associated with P1 remained intact and new intense PV anomalies continually formed, dissipated, or merged with this PV maximum so that a PV feature tracking with P1 was present through the end of the simulation. It is for this reason that we continue to label this feature P1. P2 moved southward to the west of P1 during this time. By 0800 UTC 24 July (Fig. 12e), convective activity was nearing its peak and multiple convective-scale PV anomalies had formed. P1 remained weak but associated with a pressure minimum. P2 moved southward and was collocated with a second pressure minimum. A third intense PV anomaly (labeled P3) formed a third pressure minimum. By 1200 UTC (Fig. 12f), P2 and P3 had moved very close to each other and merged by 1800 UTC (Fig. 12g). After this time, similar to P1, the pressure minimum associated with the merged P2–P3 PV maximum (labeled P2–3 in Figs. 12g,h) remained intact, with multiple convective-scale PV anomalies forming, merging, and growing following the two distinct pressure minima (i.e., P1 and P2–3).

The formation and coalescence of the convectively generated PV anomalies into mesoscale PV features and their relationship to the low-level circulation are illustrated in Fig. 13. Beginning at 0600 UTC 24 July (Fig. 13a), convection had just become more intense and widespread and the three PV features, P1, P2, and P3, were apparent in the PV fields. The low-level flow had just formed a closed cyclonic circulation, with the PV anomalies along the inner edge of the stronger cyclonic flow. By 1200 UTC (Fig. 13b), the number of PV anomalies had increased and mergers had begun. The circulation was now centered on anomalies P2 and P3. Over the next 6 h (Fig. 13c), convective-scale PV anomalies continued to form and coalesce into two growing regions of high PV, both appearing to contribute equally to the low-level circulation. Finally, by the end of the simulation, 0000 UTC 25 July (Fig. 13d), the PV had coalesced into two very distinct mesoscale regions of high PV, with the circulation center collocated with P1.

At midlevels (500 hPa; Fig. 14), a similar evolution occurred. At 0600 UTC 24 July (Fig. 14a), several intense PV anomalies were dispersed across the region. The flow showed strong cyclonic curvature but was not closed at this time in a ground-relative reference frame (a closed circulation in a wave-relative reference frame is centered near 21.6°N, 95.2°W). By 1200 UTC (Fig. 14b), the number of intense PV features increased, with some features already indicating mergers, similar to the situation at lower levels at this time. Six hours later (Fig. 14c), a significant merger of PV occurred and a strong closed cyclonic circulation had formed. This pattern was maintained for the remaining 6 h of the simulation. As implied by Fig. 14, both convective and stratiform processes played a likely role in this intensification and consolidation of PV, with stratiform processes dominating after 1500 UTC 24 July.

\(^4\) The PV fields were smoothed using 10 passes of a 9-point-weighted smoother [Eq. (11–107) of Haltiner and Williams 1980] to make the plots more legible, so PV features appear larger than their original unfiltered size.

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**Fig. 11.** Time series of vertical profiles of the area-averaged potential vorticity (0.2-PVU intervals) in (a) convective and (b) stratiform regions for the Control simulation. Negative values are lightly shaded. Positive values ≥0.6 PVU are darkly shaded.
The Doppler observations reveal vortical features resembling those seen in the simulations at low levels. At 0330 UTC (Fig. 15a), a strong localized vortex at 2-km altitude was collocated with an intense convective cell (Fig. 6b; 20.25°N, 95.4°W) on the northern or northwestern side of the cyclonic circulation, with a sharp turning in the wind field similar to that seen in Fig. 13a. A second intense vortex was at 2.5 and 3 km about 20 km to the south and was also coincident with a sharp cyclonic turning of the winds. The two vortices are qualitatively similar to P3 and P2 in Fig. 13a but are separated by about half the distance. Doppler data at 2130 UTC (Fig. 15b) also reveal multiple small-scale vortices embedded within the broader cyclonic flow. While their structure is different from that seen in the simulation at this time, the radar analyses do demonstrate that these small-scale vortices were regular features of the flow near the developing storm center.
5. The Sim2 simulation

The simulation initialized at 0600 UTC 22 July produced a storm evolution that best compared to the available observations (QuikSCAT winds, TRMM precipitation, Doppler radar, dropsondes), although it led to sea level pressures that were generally weaker than observed. Because of the weak background vortex, the merger of convectively generated PV anomalies occurred relatively slowly and primarily in the last 12 h of the simulation. In the interest of knowing how the evolution might change if the background vortex were stronger, a simulation initialized at 1200 UTC 21 July is described in this section because it produced a stronger vortex when the system was east of the Yucatan Peninsula. While the evolution of the storm is qualitatively similar to the Control run, there are some key differences that will be highlighted here.

The evolution of the simulated reflectivity and winds at 500 m is shown in Fig. 16. By 1200 UTC 22 July (Fig. 16a), a prominent vortex formed in association with convection on the coast of Belize while another region of convection extended eastward from the northeastern coast of the Yucatan. This pattern is similar to the TRMM image near this time and the location of low to midlevel cyclonic circulation (Fig. 3a). By midday on July 23 (Fig. 16b), the
system had crossed over the Yucatan and entered the Bay of Campeche. The circulation was located farther north and appeared to be stronger than that indicated by QuikSCAT (Fig. 5a), and the convection was more active than that suggested by GOES satellite imagery (not shown). By 0600 UTC 24 July (Fig. 16c), stronger convection had developed, consistent with the observations and with the Control simulation. The storm continued its movement northwestward (Fig. 16d), making landfall around 0000 UTC 25 July about 1.4° latitude too far north.

The formation of the initial vortex and its subsequent evolution are examined, as in Fig. 9, through time series of the vertical profiles of various quantities within a radius of 300 km. The area-averaged vertical motion (Fig. 17a) shows the development of very strong convection during the first 24 h of the simulation in association with the systems on the eastern side of the Yucatan Peninsula. A breakdown of this vertical motion into its convective and stratiform components (Figs. 18a,b) indicates strong and deep convection through the first 15 h of simulation, with shallower or fewer deep cells thereafter through midday on July 23. Stratiform vertical motion developed quickly and peaked just after 0000 UTC 22 July (Fig. 18b), gradually diminishing by 0000 UTC 23 July as deep convection over the broader region subsided, although it continued locally near the storm core.

As with the Control simulation, identification of a center location was difficult until about 0600 UTC July 22.

**FIG. 13.** PV and vector winds at 850 hPa at (a) 0600, (b) 1200, and (c) 1800 UTC 24 Jul and at (d) 0000 UTC 25 Jul. PV shading thresholds are at 1 and 3 PVU. The thin solid line indicates the coastline. Features labeled P1–P3 are described in the text.
As a result, the center location was determined subjectively based initially on the convection at earliest stages and later on the 850-hPa geopotential heights as the vortex developed. Consequently, during the first 18 h of simulation the tangential and radial velocity fields must be viewed with caution. The results in Fig. 17c suggest very deep inflow and convergence up to ~400 hPa, with strong divergence above, consistent with the vertical motion evolution (Fig. 17a). Although the tangential velocities prior to 0600 UTC 22 July contain higher errors as a result of uncertainties in the center position during early stages of the simulation, the tangential velocities are consistent with the deep layer of convergence, showing the rapid development of deep cyclonic flow by 1200 UTC 22 July (see also the PV field in Fig. 17f). With the weakening of convection on 23 July, the vortex became somewhat weaker and shallower and the inflow was confined mostly to very low levels. The development on 24 July was very similar to the Control case, with slow strengthening and deepening of the vortex, peak inflow at lower levels, a second peak at mid to upper levels, and strong outflow at upper levels.

The relative humidity field (Fig. 17d) shows a deep layer of saturation during the first 24 h, a period of mid-to-upper-level drying during 23 July when there was stronger mean subsidence in the environment, and then a return to nearly saturated conditions on 24 July. The evolution of the warm anomaly in Fig. 17e shows strong mid- to upper-level warming and low-level cooling during the first 24 h, followed by a gradual lowering of the warm

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**Fig. 14.** As in Fig. 13, but for 500 hPa.
Fig. 15. Dual-Doppler-derived relative vorticity for (a) 0330 and (b) 2130 UTC 24 Jul. The main panels in (a) and (b) show the vorticity (shading) and winds at 2-km altitude. Smaller boxes show vorticity and winds at 1.5, 2.5, and 3 km for the area indicated by the dashed rectangles.
anomaly and reduction of the depth of the cool air at low levels during 23 July. With redevelopment of convection on 24 July, a double-peaked structure emerged in the temperature anomaly field, with maximum warm anomalies at lower and upper levels near the levels of peak upward motion in convective regions (Fig. 18a).

The PV field (Fig. 17f) shows the development of a deep layer of PV that is maximum around 600–500 hPa around 1800 UTC 22 July. When the PV is examined separately in convective and stratiform regions (Figs. 18c,d), the results again show significant PV at low levels in convective regions and stronger PV at midlevels in stratiform regions. Some of the high PV above 600 hPa in convective regions between 0000 UTC 22 July and 0000 UTC 23 July may result from convection developing within stratiform areas already possessing high midlevel
PV, while stratiform areas with high low-level PV may contain some PV originally associated with convection. In general, though, convective regions play a large role in generating low-level PV while stratiform regions enhance midlevel PV.

The relationship between the PV anomalies and the flow at 850 and 500 hPa are shown in Figs. 19 and 20, respectively. At 850 hPa, scattered and isolated PV anomalies quickly merged to form a larger, more intense PV anomaly at 1200 UTC 22 July at the Belize coast (Fig. 19a), with a strong cyclonic circulation. Animations of PV show that this strong PV anomaly persisted for ~24 h, maintaining itself by merging with the nearby weaker convectively generated PV. Its presence and its merger with these convectively generated PV anomalies are evident in the high area-averaged PV seen in Fig. 18c. By 1200 UTC 23 July (Fig. 19b), the low-level PV was characterized by a few areas of intense PV associated with convection as well as with high PV located in nonprecipitating areas but generated within earlier convection. As convective activity increased around 0000 UTC 24 July (cf. Figs. 17a and 18d), new convective-scale regions of high PV were forming and merging with pre-existing high PV. Over the next 18 h (Figs. 19c,d), the merger of these convectively generated PV anomalies led to the gradual intensification and growth of the vortex.

At 500 hPa, early stages (Fig. 20a) were characterized by widely scattered regions of high PV located primarily in stratiform and nonprecipitating anvil regions, with smaller contributions from convection. At 1200 UTC 23 July (Fig. 20b), with the decrease in both convective and stratiform vertical motions (Figs. 18a,b), the number and area of intense PV features decreased. The flow was predominantly southerly to the east of the high-PV region and easterly within the high-PV region of the southern Gulf of Mexico, with no closed circulation found within the domain. As convection increased by 0000 UTC 24 July and continued through the end of the simulation (Figs. 18, 20c, and 20d), new regions of high PV formed within both convective and stratiform areas, gradually merging after 0600 UTC 24 July to form a mesoscale region of high PV. At 0000 UTC 24 July, the midlevel flow was predominantly southerly to southeasterly within and to the east of the high-PV region, but after the rapid mergers of PV after 0600 UTC (Figs. 20c,d) the flow very rapidly developed a closed circulation. During 24 July, the 500-hPa high PV was generally located north and west of the 850-hPa high PV, suggesting that the cyclonic
flow at 850 hPa was primarily associated with the 850-hPa PV rather than a downward projection of the midlevel PV. Since the 850-hPa high PV was generated primarily by deep convection, the results suggest that VHTs played a key, if not primary, role in the spinup of the low-level flow.

6. Discussion

Several hypotheses have been put forward for the mechanism(s) of generation of surface circulation associated with tropical cyclones. This study makes use of results from two numerical simulations of the genesis of Tropical Storm Gert (2005) to investigate the development of low-level circulation and its relationship to the precipitation evolution. The roles of convective and stratiform precipitation processes within the mesoscale precipitation system that formed Gert are discussed.

In Bister and Emanuel’s (1997) conceptual model, the precursor to genesis (development of the surface vortex) is a mesoscale region of stratiform precipitation with relatively dry conditions at lower to midlevels in a mesoscale downdraft. The vortex lowers to the surface as evaporation of the precipitation gradually moistens the lower layers and the peak in the evaporative cooling profile nears the surface. Once the lower layer is moistened, cold downdrafts are decreased or eliminated, surface latent and sensible heat fluxes increase, and subsequent convection readily increases the low-level circulation. The nearly constant high values of relative humidity below 500 hPa and the lack of a long-lived stratiform precursor suggest that the Bister and Emanuel mechanism was not a factor in the development of Gert. If the Bister and Emanuel process played a role in the development of Gert, it must have done so prior to the initial times of the simulations.

Ritchie and Holland (1997) and Simpson et al. (1997) suggested that merger of midlevel mesoscale vortices associated with the stratiform precipitation regions of multiple MCSs can enlarge the scale of the merged vortex in both the horizontal and vertical extent, eventually leading to formation of a surface circulation. While undoubtedly such a process can play a role in some events, in neither the Gert observations nor the simulations is there evidence of merger of midlevel mesoscale vortices induced by multiple MCSs. Instead, the merger process is associated with both low- and midlevel smaller-scale vortices within an MCS, with convective-scale processes playing a major role in PV concentration and merger.

Nolan (2007) used high-resolution WRF simulations initialized with idealized vortices to examine triggers for tropical cyclogenesis. He found that the inner-core region becomes humidified by moist detrainment from deep convection and once the core relative humidity exceeds 80% over most of the depth of the troposphere, the midlevel vortex contracts and intensifies. When the midlevel vortex reaches sufficient strength and the inner core is nearly saturated, a smaller-scale vortex forms very rapidly near the surface in association with a VHT and becomes the core of an intensifying cyclone. In the Gert simulation, the onset of intensification on 23 July occurs with the development of deep convection on that day. Prior to this convection, relative humidity above ~500 hPa is at a minimum midday on 23 July. Relative

![Figure 18](image-url)
humidity above 500 mb increases rapidly with the onset of deep convection in the later part of 23 July, thus meeting one of the requirements described by Nolan (2007). However, in the simulated Gert, intensification of the low-level vortex is concurrent with, if not prior to, intensification of the midlevel vortex and the storm never reaches a point of rapid intensification as seen in Nolan’s idealized cases. The differences between the evolution of Gert and Nolan’s idealized simulations may lie in some aspect of the more complicated environment of Gert that is not included in the idealized initial environment but that more readily facilitates spinup at low levels.

Vortex development would likely accelerate as the frequency of convective updrafts increases with time (Nolan 2007) and as the peak vertical motion lowers with time because of stabilization of the environment (Raymond and Sessions 2007). However, at least in terms of the area-averaged vertical motion, the profile becomes increasingly stratiform (Fig. 9a) during the latter part of 24 July and the convective profile shows no lowering with time (Fig. 10a). This result may explain, in part, the very slow development of Gert into a tropical storm.

The evolution of the low-level potential vorticity field in the Gert simulations is qualitatively similar to that seen in idealized simulations by Van Sang et al. (2008). In their experiments, storm intensification begins with the development of a ring of convection that produces

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**Fig. 19.** PV and vector winds at 850 hPa for the Sim2 simulation at (a) 1200 UTC 22 Jul, (b) 1200 UTC 23 Jul, (c) 0600 UTC 24 Jul, and (d) 1800 UTC 24 Jul. PV shading thresholds are at 1 and 3 PVU.
intense small-scale vorticity dipoles, with strong cyclonic vorticity and weak anticyclonic vorticity. Over time, the cyclonic vorticity anomalies merge and the anticyclonic vorticity becomes axisymmetrized. As a result, the VHTs contribute directly to the storm-scale spinup. The evolution of the 850-hPa PV field shown in Figs. 12e–h shows a similar pattern of gradual concentration of cyclonic PV in the inner region of the storm, with a greater prevalence of negative PV farther out from the center, particularly after convection develops on 24 July. These results confirm the findings of Hendricks et al. (2004), Montgomery et al. (2006), Van Sang et al. (2008), and Shin and Smith (2008) that VHTs play a key, if not a leading, role in intensifying low-level circulation during tropical cyclogenesis.

7. Conclusions

This paper examined high-resolution simulations of Tropical Storm Gert (2005), which formed in the Gulf of Mexico during NASA’s Tropical Cloud Systems and Processes Experiment. Simulations were conducted using the Weather Research and Forecasting numerical prediction model and results were thoroughly validated against satellite and airborne datasets. Two simulations were examined: one that better matches available observations but underpredicts the storm’s minimum sea level pressure and a second one that somewhat overintensifies the storm but provides a set of simulations that encapsulates the overall genesis and development characteristics of the observed storm.
A convective–stratiform precipitation separation technique was applied to investigate the roles of convective and stratiform precipitation processes in the development of Gert. As is typical for precipitation systems (Houze 1993), system evolution was characterized by intense and deep convective upward motions followed by increasing stratiform-type vertical motions (upper-level ascent, low-level descent). Potential vorticity in convective regions was strongest at low levels, but with high PV extending up to almost 300 hPa. Stratiform region PV was strongest at midlevels. Given the evolution of convective and stratiform regions mentioned above, this result suggests that convective processes act to spin up lower levels prior to the spinup of middle levels by stratiform processes. After convection subsides, stratiform processes continue to spin up middle levels for some period of time. Subsequent convective systems occurring in the higher-PV wake of the previous one would act also to enhance PV at low levels via convection prior to enhancing midlevel PV via stratiform processes. This process was seen in the case of Gert, with the two primary episodes of convective system development on 22 July and late on 23 July into the 24th.

Intense VHTs were prominent features of the low-level cyclonic vorticity field. The most prominent PV anomalies persisted more than 6 h and often were associated with localized minima in the sea level pressure field. A gradual aggregation of the PV occurred, with cyclonic PV becoming more concentrated near the storm center. In the case of the weaker storm development (the Control experiment), two intense PV regions dominated the flow, forming two storm centers, each gradually intensifying as they merged with newer VHTs. In the case of the stronger surface pressure development (the Sim2 experiment), the VHTs merged into a single low pressure center, gradually increasing the mean vorticity near the center. As pointed out by Montgomery et al. (2006), not only do these hot towers act to locally increase the vorticity, they also contribute to the evolution of the system-scale mean secondary circulation, increasing the low and midlevel inflow and converging the background cyclonic vorticity and the convective-scale cyclonic vorticity generated by the hot towers.

Nearly concurrently with this VHT-induced development, stratiform precipitation processes strongly enhanced the mean inflow and convergence at middle levels, rapidly increasing the midlevel vorticity. However, the stratiform vertical motion profile is such that while it increases midlevel vorticity, it decreases vorticity near the surface as a result of low-level divergence. Consequently, the results presented here for Gert are in agreement with Tory et al. (2006a,b) in that while stratiform precipitation regions may significantly increase cyclonic circulation at midlevels, convective vortex enhancement at low to midlevels is likely necessary for genesis.

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