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ABSTRACT

Airborne Doppler radar observations are used to document the structure of three miniature supercells embedded in an outer rainband of Hurricane Ivan on 15 September 2004. The cells were located more than 100 km offshore, beyond the Doppler range of coastal radars. The combination of large CAPE, large vertical wind shear, and moderate cell-relative helicity with an apparent midlevel dry air intrusion provided an offshore environment supporting rotating storms. Each shallow cell contained a 5–7-km-diameter mesocyclonic updraft with midlevel updraft and vorticity maxima that exceeded 6 m s\(^{-1}\) and 0.008 s\(^{-1}\), respectively. Such offshore structures are consistent with miniature supercells observed onshore in association with tropical cyclone tornado outbreaks. The strong updrafts resulted from a combination of kinematic convergence, thermal instability, and shear-induced vertical perturbation pressure gradients. Mesocyclone production largely resulted from the tilting and subsequent stretching of low-level horizontal streamwise vorticity into the vertical by the strong updrafts. Evidence of baroclinic contributions from inflow along cell-generated outflow boundaries was minimal. The miniature supercells persisted for at least 3 h during transit from offshore to onshore. Tornadoes were reported in association with two cells soon after moving onshore. These observations build upon a growing body of evidence suggesting that miniature supercells often develop offshore in the outer rainbands of tropical cyclones.

1. Introduction

Supercells are long-lived convective storms that contain a mesocyclonic updraft and downdraft while often exhibiting a distinctive “hook echo” in radar reflectivity imagery. In the midlatitudes, the majority of large and violent tornadoes are spawned by supercells (e.g., Davies-Jones et al. 2001). In landfalling tropical cyclones, most tornadoes are spawned by “miniature” supercells (Spratt et al. 1997; Suzuki et al. 2000; McCaul et al. 2004). Although they are often shallower, less intense, and shorter-lived than their midlatitude counterparts, such supercells have been responsible for multiple deadly tornado outbreaks (McCaul 1991; Hagemeyer 1997). However, our understanding and ability to forecast miniature supercells and tornadoes in tropical cyclones remains limited, in part, because available observation networks are unable to regularly monitor the relevant offshore environment and cell evolution.

Numerous studies have documented common environmental characteristics associated with tornadic miniature supercells observed in tropical cyclones (e.g., Novlan and Gray 1974; McCaul 1991; Curtis 2004; Edwards and Pietrycha 2006). The majority of events occurred in outer rainbands that were embedded within the onshore flow of the northeast quadrant (often the right-front quadrant relative to storm motion). Favorable environments often contained 1) moderate low-level storm-relative helicity, 2) moderate CAPE with maximum buoyancy below 500 hPa, 3) moist low levels with a low lifting condensation level, 4) relatively dry air at midlevels, and 5) a preexisting low-level boundary (such as a convergence axis or baroclinic front). The numerical simulations of McCaul and Weisman (1996, 2001) further demonstrated that the formation of miniature supercells was favored in environments with the buoyancy and shear maxima concentrated in the lower troposphere.

Previous studies also have documented the structure and evolution of miniature supercells within landfalling tropical cyclones using Doppler radar observations (e.g., Spratt et al. 1997; McCaul et al. 2004). The tornado-producing cells typically were characterized by >50-dBZ echoes, low-level inflow notches, and echo tops <10 km.
Their mesocyclones were shallow (~4 km) with small diameters (~2 km) and mean shear vorticity magnitudes of ~0.005 s$^{-1}$ that could be tracked for 1–2 h. The mesocyclones were often identified ~30 min prior to reported tornado production.

Although the vast majority of supercells have been documented over land as a result of radar limitations offshore, there is a growing body of evidence suggesting that miniature supercells in tropical cyclones often develop well offshore. Indeed, prior studies have demonstrated that offshore environments can be conducive to supercell formation (Bogner et al. 2000; Baker et al. 2009), documented offshore supercells using coastal Doppler radars (Spratt et al. 1997; Rao et al. 2005; Lee et al. 2008), and noted reports of waterspouts and tornadoes within a few kilometers of the coastline (Gentry 1983; Hagemeyer 1997). This paper contributes to such evidence by providing the first detailed observations of tropical cyclone miniature supercells located more than 100 km offshore and beyond the Doppler range of operational coastal radars.

The objective of this study is to document the three-dimensional structure and evolution of three distinct miniature supercells embedded within an offshore outer rainband of Hurricane Ivan (2004) using primarily airborne Doppler radar observations. Our specific goals include the following:

1) Describe the offshore environment supporting the supercells.

2) Document the three-dimensional structure and airflow of the offshore supercells.

3) Determine the physical processes responsible for supercell formation and maintenance.

4) Describe the supercells’ subsequent evolution and coastline tornadogenesis.

We hope that this study will enhance our understanding of tropical cyclone convection and provide further insight as to how tornadoes develop within landfalling systems.


Hurricane Ivan formed in the central Atlantic on 5 September 2004 from a strong African easterly wave. On 9 September, Ivan moved into the Caribbean Sea as a major hurricane. On 12 September, the hurricane achieved a maximum intensity of 75 m s$^{-1}$ with a central pressure of 910 hPa as it passed south of Grand Cayman. Soon afterward, Ivan began to move northward in response to a weakening in the subtropical ridge. After passing over the western tip of Cuba, the system entered the Gulf of Mexico on 14 September as a strong category-4 hurricane. Ivan made landfall near Gulf Shores, Alabama, at approximately 0650 UTC 16 September with maximum sustained winds of 54 m s$^{-1}$ and a central pressure of 943 hPa. Over the next 36 h, the system moved toward the northeast and weakened significantly as it crossed the southeastern United States. Franklin et al. (2006) provides a detailed account of Ivan’s history.

The observations analyzed in this study were collected between 1400 and 2100 UTC 15 September when Ivan was located ~250 km offshore in the northern Gulf of Mexico (Fig. 1). During this period, Ivan was moving northward at ~5.7 m s$^{-1}$ in response to a strong subtropical ridge centered east of the Bahamas and an approaching midlatitude trough extending southward into Texas. The storm was weakening slowly (maximum winds decreased from 59 to 56 m s$^{-1}$) in part because of increasing vertical shear, decreasing sea surface temperatures, and the ingestion of dry air into the core. At the same time, a prominent outer rainband developed ~350 km east of the storm center and extended through the northeast (or right front) quadrant. The band was located just east (or outside) of an apparent dry air intrusion that wrapped through the southern and eastern quadrants. Examination of animated satellite and land-based radar imagery revealed that the band was composed of multiple long-lived convective cells that initially formed southeast of the storm center, moved along the band though the eastern quadrant, and then moved onshore into the Florida Panhandle. This paper examines a subset of these cells.

Ivan produced at least 122 tornadoes as the storm crossed the southeast United States [National Climatic Data Center (NCDC 2004)], which is currently the record for a tropical cyclone. A period of significant tornado activity occurred on the afternoon and evening of 15 September as Ivan approached the Gulf Coast. During this time, the cells embedded within the prominent outer rainband spawned 17 reported tornadoes (5 produced F1 damage and 2 produced F2 damage) across portions of the Florida Panhandle and southeast Alabama. These tornadoes were responsible for 6 deaths and 13 injuries. A cell examined herein spawned one of these deadly tornadoes.

3. Data and analysis methods

On 15 September 2004, the National Oceanic and Atmospheric Administration (NOAA) Hurricane Research Division conducted a multiplane investigation of Hurricane Ivan. Although the primary goals of the research mission were unrelated to the prominent outer rainband east of the storm center (see Fig. 1), considerable data were collected in and near the rainband from
a variety of platforms. The dataset employed here primarily consists of airborne Doppler radar observations used to document the rainband’s offshore convective cell structure. The data were supplemented with flight-level, dropsonde, rawinsonde, and surface observations to analyze the rainband environment and cell structure. Observations from the Tallahassee (TLH) and Tampa Bay (TBW), Florida, operational Weather Surveillance Radar-1988 Doppler (WSR-88D) were used to provide an overview of the offshore rainband structure, track individual cells, and document the structural evolution of individual cells as they moved onshore. Figure 2 provides a geographic overview of these data with respect to the rainband at the time of interest.

**a. Airborne radar observations**

A NOAA WP-3D aircraft crossed the rainband between 1758 and 1810 UTC at ~2.5 km AGL while en route to the eyewall. The aircraft was equipped with a 5.5-cm wavelength lower fuselage (LF) radar and a 3.2-cm wavelength tail Doppler radar. Details of the radars are described in Jorgensen (1984). During the rainband crossing the aircraft employed the fore–aft scanning technique (FAST) to collect Doppler velocity observations (Gamache et al. 1995). The FAST geometry (alternating conical scans ~20° fore and aft of the plane normal to the aircraft track) permitted a dual-Doppler analysis of the three-dimensional wind field for a portion of the rainband (see Fig. 2b). This geometry also dictates a time lag between each radar observation at a given location within the analysis domain. In the present study, the average time lag was <2 min, and thus structural details of individual convective cells should be well represented.

The Doppler analysis methodology described in Gamache (1997) and Reasor et al. (2009) was used to construct the three-dimensional wind field of the offshore rainband segment. Initially, the raw reflectivity and Doppler velocity data from the tail radar were edited following standard procedures that remove spurious echoes (e.g., sea clutter) and incorporate appropriate navigation corrections (Bosart et al. 2002). Then, the edited fields were interpolated to a band-centered 90 × 90 km Cartesian domain extending from the surface to 18 km AGL with uniform horizontal and vertical grid spacing of 1.5 and 0.5 km, respectively. The initial values at each grid point were determined through a Gaussian weighting of all nearby observations using a cutoff radius of 3 km (1 km) and an $e$-folding distance of 0.75 km (0.25 km) in the horizontal (vertical) direction. Finally, the variational methodology was used to solve simultaneously the radar projection equations and anelastic mass continuity equation for the three-dimensional wind field while incorporating an appropriate motion for the target cells. In application, Gamache (1997) found this method superior to traditional methods (e.g., Jorgensen et al. 1983), but time evolution of the wind field, beam filling issues, and variable cell motions still could make significant contributions to the error.

The quality of the Doppler analysis was assessed by comparing flight-level observations with the Doppler-derived winds at 2.5 km AGL following Gamache et al. (1995). Initially, the filtered flight-level data (see section 3b) were re navigated and interpolated to the Doppler...
analysis grid. Then, a gridpoint-by-gridpoint comparison of each Doppler and interpolated flight-level wind component was conducted. To facilitate direct comparison with previous studies, summary statistics for the zonal, meridional, tangential, radial, along-track, cross-track, and vertical wind components were computed from the 125 available grid points (see Table 1). Overall, the summary statistics were similar to values previously reported in hurricanes (Marks et al. 1992; Gamache et al. 1995; Reasor et al. 2009). Mean differences (or biases) are very small and fall within the accuracy range of the flight-level observations. The low linear correlation coefficient (0.27) for the vertical wind component resulted, in part, from the inherent spatial smoothing involved when Doppler-derived vertical motions are constrained by mass continuity and the horizontal winds (Marks et al. 1992). The relatively weak vertical motions along much of the flight path (the aircraft passed between cells) also likely contributed to the low correlation. It should be noted, however, that the individual distributions of flight-level and Doppler-derived vertical motions were very similar to those found in previous studies.

b. Flight-level observations

The NOAA WP-3D flight-level sensors are described in Jorgensen (1984). Initially, the 1-Hz data obtained

1 The aircraft ground speed was −125 m s⁻¹ during the rainband crossing.
During the rainband crossing were smoothed using a running Bartlett filter designed to preserve scales greater than 1.5 km (the Doppler grid resolution) without shifting significant peaks. The vertical velocity (w) data were analyzed following Eastin et al. (2005) to remove a small nonzero offset (0.12 m s⁻¹). The available temperature (T) and dewpoint (T_d) data were obtained by the Rosemount thermometer and General Eastern cooled mirror, respectively, which are susceptible to errors induced by sensor wetting in clouds and precipitation (e.g., Eastin et al. 2002). During wetting, the temperature sensor behaves more like a wet-bulb thermometer and reports erroneously low values, whereas the chilled mirror measures erroneously high dewpoint values that often exceed the low temperatures. Any such wetting errors were reduced following the recommendations of Zipser et al. (1981), but error magnitudes on the order of 0.5°C may still remain (Eastin et al. 2002). Equivalent potential temperatures (θ_e) were computed following Bolton (1980) and may be underestimated by ~3 K during periods of instrument wetting.

c. Sounding observations

Two soundings were selected to represent the rainband environment. The first was the operational rawinsonde launched at 1800 UTC from Tampa Bay. The second was a GPS dropsonde (Hock and Franklin 1999) deployed at 1807 UTC by the NOAA G-IV aircraft during a synoptic surveillance mission (see Fig. 2b). Their selection from the available soundings was based on proximity in time and space: they were the closest upwind soundings (within 1 h and 250 km) from the rainband crossing location. The GPS sounding was processed and quality controlled using the Atmospheric Sounding Processing Environment (ASPEN) program (available online at http://www.eol.ucar.edu/rtf/facilities/software/aspen/aspen.html) developed by NCAR. Further scrutiny revealed no common data issues, neither a loss of near-surface winds (Franklin et al. 2003) nor any questionable segments in the thermodynamic profile (Barnes 2008).

To facilitate comparison with prior studies, the following stability and vertical shear parameters were computed from each sounding: most unstable CAPE (MUCAPE; Doswell and Rasmussen 1994), mean-layer CAPE (MLCAPE) for a mean parcel from the lowest 100-m layer, surface-based CAPE (SBCAPE), surface-based convective inhibition (SBCIN), lifting condensation level (LCL), 0–3-km SBCAPE, surface θ_e, 0–1-km and 0–6-km shear vector magnitude, effective bulk shear magnitude (Thompson et al. 2007), 0–1-km and 0–3-km cell-relative helicity (CRH; Davies-Jones et al. 2001), effective CRH (Thompson et al. 2007), bulk Richardson number (BRN) and BRN shear (Weisman and Klemp 1982), 0–1-km and 0–3-km energy helicity index (EHI; Rasmussen 2003), supercell composite parameter (SCP; Thompson et al. 2004), significant tornado parameter (STP; Thompson et al. 2004), and the 10–500-m critical angle (Esterheld and Giuliano 2008).

In computing cell-relative quantities, cell motions were obtained by tracking multiple convective cells in close proximity (within 3 h and 50 km) to each sounding using animated TBW and TLH radar imagery. Estimates of cell motion were computed using the modified Bunkers et al. (2000) technique described in Ramsay and Doswell (2005), and they exhibited a small right bias (~15°–20°) compared to the motions obtained by cell tracking.

d. Surface observations

Offshore surface observations were very limited. A single moored buoy (station 42036) maintained by the National Data Buoy Center (NDBC) was located within the Doppler analysis domain (see Fig. 2b) and thus used to provide representative surface observations near the rainband. Detailed information on platform configuration, sensor descriptions and accuracy, data acquisition, and quality control is available at the NDBC Web site (available online at http://www.ndbc.noaa.gov/). Further scrutiny revealed no wet-bulb contamination as defined by Cione et al. (2000).

Surface winds were available every 10 min, based on an 8-min averaging period. For better consistency with the other observations and prior studies of the hurricane boundary layer, the winds were normalized to a maximum 1-min sustained averaging time using a typical overwater roughness length (Powell et al. 1996). The

TABLE 1. Mean differences (bias), RMS differences, and linear correlation coefficients between aircraft flight level and Doppler-derived wind components for all data points. Statistics for the zonal (U), meridional (V), vertical (W), radial (VR), tangential (VT), along-track (VA), and cross-track (VC) wind components are listed. A positive (negative) bias indicates that the Doppler-derived wind is more positive (more negative) than the flight-level wind.

<table>
<thead>
<tr>
<th>Wind component</th>
<th>U</th>
<th>V</th>
<th>W</th>
<th>VR</th>
<th>VT</th>
<th>VA</th>
<th>VC</th>
</tr>
</thead>
<tbody>
<tr>
<td>Statistic</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Bias (m s⁻¹)</td>
<td>0.17</td>
<td>−0.35</td>
<td>0.02</td>
<td>0.40</td>
<td>−0.36</td>
<td>0.18</td>
<td>0.01</td>
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<tr>
<td>RMS (m s⁻¹)</td>
<td>1.55</td>
<td>1.17</td>
<td>1.12</td>
<td>1.63</td>
<td>1.23</td>
<td>1.35</td>
<td>1.39</td>
</tr>
<tr>
<td>Correlation</td>
<td>0.93</td>
<td>0.64</td>
<td>0.27</td>
<td>0.89</td>
<td>0.77</td>
<td>0.99</td>
<td>0.99</td>
</tr>
</tbody>
</table>

Studies of midlatitude supercells regularly evaluate storm-relative parameters. Here, we have adopted the term cell relative for such evaluations, while reserving the term storm relative for those analyses with respect to the moving tropical cyclone center.
winds also were adjusted to the standard 10-m AGL reference height (from 5 m) following Liu et al. (1979). These adjustments were equivalent to applying an amplification factor of \( \frac{1}{1.2} \) to the 8-min winds. In addition, the hourly temperature and dewpoint observations were adjusted to the 10-m height, assuming a dry adiabatic lapse rate and a constant mixing ratio following Cione et al. (2000).

4. Offshore rainband and supercell structure

a. Environmental characteristics

Figures 3 and 4 present the thermodynamic profiles and hodographs representative of the rainband environment. The TBW sounding was taken \( \approx 220 \) km east-southeast of the rainband segment, whereas the GPS sounding was located \( \approx 50 \) km to the southwest near the apparent modest dry air intrusion (see Fig. 1). These soundings and the analyzed rainband segment were located in the northeast (or right front) quadrant (see Fig. 2b). Table 2 lists the instability and shear parameters computed from each sounding.

The rainband environment supported the potential for rotating convection. Both soundings exhibited a shallow moist layer near the surface (below \( \approx 1 \) km), drier air at midlevels (\( \approx 2-5 \) km), and moist air aloft (Fig. 3). Midlevel RH minima approached 40\%-50\% near \( \approx 4-5 \) km AGL. CAPE values (Table 2) and their vertical distributions were typical for the \( \approx 300-500 \)-km range from the storm center in both offshore (Bogner et al. 2000) and tornadic landfalling (McCaul 1991) hurricanes. The wind profiles (Fig. 4) exhibited south-easterly low-level winds that veered with height, resulting in a 0–6-km AGL vertical shear vector oriented toward the northeast. Shear magnitudes and cell-relative helicities (Table 2) were equivalent to or exceeded mean values observed at similar ranges in hurricanes (Novlan and Gray 1974; McCaul 1991; Bogner et al. 2000). A third wind profile was constructed by combining the 1800 UTC surface wind from the moored buoy (BY) with the Doppler-derived (DD) winds above the buoy location (DD/BY; see Fig. 4b). This profile, located \( \approx 10–20 \) km east of the strongest cells, provides a representation of the local inflow environment and exhibited remarkably similar structure (and shear parameters) to the GPS wind profile. Finally, based on midlatitude forecast criteria, the BRN (<50), SCP (>1), and 0–3 km EHI (>1) suggest the environment was conducive to supercell formation (and marginally conducive to tornadogenesis). Baker et al. (2009) should be consulted for a more detailed discussion of all available soundings and their forecast parameters.

b. Evolution of the band and cells

Figure 2 provides an overview of rainband structure and evolution between 1732 and 1831 UTC as viewed from TBW and TLH. During this period, the analyzed offshore rainband segment was beyond the Doppler
range (~175 km) of both coastal radars (~220 km from the segment), but reflectivity data\textsuperscript{3} were available. The band was composed of multiple convective cells (maximum dBZ > 40) spaced at ~20–30-km intervals. Cells south of the analyzed segment (Figs. 2a,b) exhibited no consistent structure, but cells to the north (Figs. 2c,d) tended to exhibit northeasterward elongation with maximum reflectivity gradients and hooklike appendages to the southwest, indicating supercells. The band axis (or line of strongest cells) was located along the western (or inner) edge adjacent to the apparent dry air intrusion (see Fig. 1). Stratiform precipitation dominated the eastern (or outer) side with increased areal extent to the north.

Although Fig. 2 depicts a variety of evolving cellular structures along the rainband, the aircraft’s LF imagery (not shown) indicated that the analyzed cells (labeled A, B, and C) remained relatively steady in shape and intensity from 1758 to 1810 UTC when the airborne Doppler data were collected. Tracking these cells via animated TLH and TBW imagery revealed a mean north-northwest motion (toward 344° at 24.8 m s\textsuperscript{21}). This earth-relative cell motion includes the storm motion (toward 355° at 5.7 m s\textsuperscript{21}) and a small outward band motion (toward 63° at 1.1 m s\textsuperscript{21}).

c. Horizontal structure

Figures 5 and 6 depict the reflectivity, cell-relative winds, vertical velocity, and vertical vorticity at 2 km AGL for the Doppler-analyzed rainband segment. Three distinct cells (labeled A, B, and C) were located along the western edge near the apparent (echo free) dry air intrusion. Similar to cells farther north, each occupied ~600 km\textsuperscript{2} (based on the 25-dBZ contour, as in Barnes et al. 1991), exhibited a southwest–northeast orientation, and contained reflectivity maxima greater than 40 dBZ\textsuperscript{4} with maximum gradients along the western edge. Cells A and B each exhibited a weak hook echo and inflow notch. Each cell also contained a closed cyclonic circulation (i.e., a mesocyclone) in the cell-relative Doppler-derived wind field, which was roughly collocated with the cell’s reflectivity, updraft, and vertical vorticity maxima. Distinct downdrafts were located in cell A (cell B) ~5–8 km north (south) of the primary updraft along the cell’s forward (rear) flank. Clearly, these offshore cells in Ivan’s outer rainband exhibited signature characteristics of supercells (e.g., Lemon and Doswell 1979).

At 2 km AGL, each primary updraft was ~6 km in diameter (based on the 2 m s\textsuperscript{21} contour) with a maximum greater than 6 m s\textsuperscript{21}. Such characteristics are representative of the largest and strongest updrafts

\textsuperscript{3} At such range, the ability of the radar reflectivity data to resolve low-level convective structure (such as inflow notches) is somewhat limited by the 1° beamwidth and the base-scan elevation. Spratt et al. (1997) provide a more detailed discussion of the long-range sampling limitations of the WSR-88D radars with specific regard to miniature supercells.

\textsuperscript{4} Individual sweeps from the tail radar exhibited reflectivity maxima of >50 dBZ at 3–4 km AGL, consistent with Spratt et al. (1997) and McCaul et al. (2004).
observed at roughly the same altitude in hurricane inner-core rainbands (e.g., Black et al. 1996; Eastin et al. 2005) and other tropical oceanic convection (e.g., Lucas et al. 1994). The collocated mesocyclones were slightly smaller ($\leq 5$ km in diameter based on the 2 s$^{-1}$ vorticity contour$^5$) with vorticity maxima greater than $8 \times 10^{-3}$ s$^{-1}$. These characteristics are consistent with those of miniature supercells observed in onshore tropical cyclone rainbands (e.g., Spratt et al. 1997; McCaul et al. 2004; Schneider and Sharp 2007).

At 5 km AGL (Figs. 7, 8), each cell’s reflectivity and updraft maxima were located northeast of their respective low-level positions, consistent with a tilt induced by the ambient vertical shear (see Fig. 4). Cell-relative flow approached from the southwest (i.e., from the apparent dry air intrusion) and penetrated across the rainband axis while diverging around each primary updraft. Such injections of drier midlevel air would support evaporative cooling, and indeed each cell showed evidence of forward-flank and rear-flank downdrafts (FFDs and RFDs, respectively). As discussed next, evidence of any mesocyclones at this altitude was minimal because the mesocyclones were shallow features.

d. Vertical structure

Figures 9–11 show representative cross sections of reflectivity, vertical velocity, vertical vorticity, and in-plane cell-relative flow through the mesocyclonic updrafts (see Fig. 6) of each miniature supercell. Cell A extended up to 8 km (based on the 25-dBZ contour) and exhibited a shallow weak-echo vault (Fig. 9). Cell-relative inflow approached from the east and was largely confined below 2 km. The primary updraft extended from the boundary layer up to $\sim 7$ km and achieved a maximum of $\sim 10$ m s$^{-1}$ in the 3–4-km layer before detraining to the northeast. The collocated mesocyclone extended up to $\sim 6$ km and exhibited distinct low-level and midlevel vorticity maxima ($\sim 8 \times 10^{-3}$ s$^{-1}$ at 1.5 km and $\sim 12 \times 10^{-3}$ s$^{-1}$ at 3.5 km, respectively). Evidence of a weak RFD ($\sim 2$ m s$^{-1}$) was confined to the 4–5-km layer at the point where the southwesterly flow from the apparent dry air intrusion collided with the rotating updraft. The more pronounced FFD (with magnitudes of 3–4 m s$^{-1}$) was associated with the midlevel evaporation.

5 Studies of midlatitude supercells traditionally have used vertical vorticity of $>10^{-2}$ s$^{-1}$ as the threshold definition for a mesocyclone (e.g., Brandes 1984). However, given the diminutive character of miniature supercells (and thus the potential undersampling of their mesocyclone’s peak radial velocities), such an arbitrary threshold may not be appropriate. The threshold used here roughly corresponds with the outer edge of the cell-relative closed cyclonic circulation.

### Table 2. Environmental stability, wind shear, and forecast parameters for the three soundings obtained near the rainband and miniature supercells. The DD/BY sounding (winds only) was obtained by combining the Doppler-derived winds with the surface wind observed at the buoy location (cf. Fig. 2b). All cell-relative parameters were computed using observed cell motions obtained by tracking proximity cells with radar. Cell motions computed following the modified Bunkers technique (Ramsay and Doswell 2005) are only listed for comparison purposes.

<table>
<thead>
<tr>
<th>Parameter</th>
<th>GPS-1807</th>
<th>DD/BY-1800</th>
<th>TBW-1800</th>
</tr>
</thead>
<tbody>
<tr>
<td>Observed cell motion (m s$^{-1}$)</td>
<td>24.8/164</td>
<td>24.8/164</td>
<td>15.8/175</td>
</tr>
<tr>
<td>Modified Bunkers cell motion (m s$^{-1}$)</td>
<td>24.3/181</td>
<td>24.3/178</td>
<td>15.1/194</td>
</tr>
<tr>
<td>MUCAPE (J kg$^{-1}$)</td>
<td>960</td>
<td>2924</td>
<td>2924</td>
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<tr>
<td>MLCAPE (J kg$^{-1}$)</td>
<td>618</td>
<td>2344</td>
<td>2344</td>
</tr>
<tr>
<td>SBCAPE (J kg$^{-1}$)</td>
<td>960</td>
<td>2924</td>
<td>2924</td>
</tr>
<tr>
<td>0–3-km SBCAPE (J kg$^{-1}$)</td>
<td>108</td>
<td>158</td>
<td>158</td>
</tr>
<tr>
<td>SBCIN (J kg$^{-1}$)</td>
<td>$-16$</td>
<td>$-12$</td>
<td>$-12$</td>
</tr>
<tr>
<td>Height of the LCL (km)</td>
<td>520</td>
<td>640</td>
<td>640</td>
</tr>
<tr>
<td>Surface $\theta_e$ (K)</td>
<td>351</td>
<td>359</td>
<td>359</td>
</tr>
<tr>
<td>0–1-km shear (m s$^{-1}$)</td>
<td>15.4/190</td>
<td>15.3/185</td>
<td>6.3/165</td>
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<td>0–6-km shear (m s$^{-1}$)</td>
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<td>28.4/215</td>
<td>17.5/205</td>
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<tr>
<td>Effective bulk shear (m s$^{-1}$)</td>
<td>21.3/211</td>
<td>16.3/208</td>
<td>16.3/208</td>
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<td>BRN shear (m$^2$ s$^{-2}$)</td>
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<td>98</td>
<td>62</td>
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<td>0–1-km CRH (m$^2$ s$^{-2}$)</td>
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<td>181</td>
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<tr>
<td>0–3-km CRH (m$^2$ s$^{-2}$)</td>
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<td>Effective CRH (m$^2$ s$^{-2}$)</td>
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<td>10–500-m critical angle ($^\circ$)</td>
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<tr>
<td>BRN SCP</td>
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<td>38</td>
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<tr>
<td>SCP</td>
<td>4.2</td>
<td>4.1</td>
<td>4.1</td>
</tr>
<tr>
<td>STP</td>
<td>0.6</td>
<td>0.7</td>
<td>0.7</td>
</tr>
<tr>
<td>0–3-km EHI (using MLCAPE)</td>
<td>0.9</td>
<td>1.4</td>
<td>1.4</td>
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<tr>
<td>0–1-km EHI (using MLCAPE)</td>
<td>0.7</td>
<td>0.7</td>
<td>0.7</td>
</tr>
</tbody>
</table>
precipitation cascade. Neither downdraft penetrated below 0.5 km.

Cell B was shallower and less intense (Fig. 10). Cell-relative inflow approached from the northeast and was confined below 1.5 km. The primary updraft extended up to \( \approx 6.5 \) km and was composed of three distinct updraft maxima. The strongest achieved \( \approx 6 \text{ m s}^{-1} \) at 2.5 km AGL and was collocated with the mesocyclone (which exhibited maximum vorticity of \( \approx 11 \times 10^{-3} \text{ s}^{-1} \) at 2 km). The lack of a weak-echo vault is consistent with the less intense updraft. A modest RFD was present and extended below 0.5 km, but no appreciable FFD was evident.

Cell C contained the strongest updraft but the weakest mesocyclone (Fig. 11). Low-level cell-relative inflow was primarily from the east-southeast and thus likely was impacted by the small transient cell located \( \approx 10 \) km upwind (see Figs. 5–8). The updraft extended from a shallow weak-echo region up to \( \approx 9 \) km, attaining a maximum of 11 m s\(^{-1}\) at 5 km. The collocated mesocyclone extended up to \( \approx 6 \) km with distinct vorticity maxima of \( \approx 7 \times 10^{-3} \text{ s}^{-1} \) at \( \approx 2 \) and \( \approx 4 \) km. Neither the RFD nor the FFD (not shown) penetrated below 1.0 km.

Clearly, several notable similarities existed among these miniature supercells. Their overall diminutive character (cf. classic midlatitude supercells) was consistent
with the local thermodynamic profile (Fig. 3a); moderate CAPE contributed to modest updraft magnitudes, whereas the vertical confinement of CAPE by the hurricane’s upper-level warm core limited updraft and mesocyclone depth (McCaul and Weisman 1996; Bogner et al. 2000). The combination of well-developed updrafts and mesocyclones with only a few weak downdrafts suggests that each cell was in the developing stage of its life cycle. Finally, the presence of multiple updraft maxima located ~0.5 km above their associated vertical vorticity maxima suggests that convergence and stretching of vertical vorticity played a role in mesocyclone formation. However, as discussed in the next section, the tilting of low-level horizontal vorticity also contributed.

5. Forces responsible for the supercells

a. Updraft production

Current conceptual models of supercell evolution suggest that persistent updraft production is supported through a combination of mesoscale convergence and local thermodynamic and dynamic forcing (e.g., Davies-Jones et al. 2001). The thermodynamic forcing results from lifting air parcels in an environment with nonzero
CAPE until they become locally buoyant and accelerate upward. The dynamic forcing results primarily from vertical perturbation pressure gradients, induced by an updraft interacting with a veering environmental vertical shear of the horizontal wind, that support upward accelerations on the updraft’s right flank relative to the shear vector (e.g., Rotunno and Klemp 1982). As noted by McCaul and Weisman (1996, 2001), environments with both maximum buoyancy and vertical shear confined to the low and midlevels are optimal for miniature supercell production via such combined forcing.

Consistent with this conceptual model, a combination of thermodynamic and shear-induced forcing appears sufficient to explain the modest low- to midlevel updrafts observed in Ivan’s supercells. Both proximity soundings exhibit ample CAPE primarily confined to low and midlevels (Fig. 3 and Table 2). Even weak near-surface convergence along the rainband axis will provide adequate lift to overcome the minimal convective inhibition (SBCIN) and elevate parcels to the local level of free convection. However, close inspection of the soundings reveals that CAPE is limited below 2 km, suggesting that the low-level updrafts may be enhanced primarily by the shear-induced vertical pressure gradients. Both the GPS and DD/BY wind profiles depict semicircular cell-relative hodographs composed of veering shear vectors that maximize in the lowest 3 km (Fig. 4). Thus, following Rotunno and Klemp (1982), persistent shear-induced updraft production would maximize on the southeast flank of each supercell’s primary updraft. It is interesting to speculate that the shallow low-level updraft maxima just east of cell B’s primary updraft (see Figs. 6, 10a) may be a manifestation of this shear-induced forcing.
b. Vorticity production

The high-resolution Doppler winds provide an opportunity to evaluate the three-dimensional vorticity field near each miniature supercell with a focus on vertical vorticity production conducive to mesocyclone development and maintenance. In Cartesian coordinates, the three-dimensional relative vorticity ($\omega$) is defined as

$$\omega = \xi \mathbf{i} + \eta \mathbf{j} + \zeta \mathbf{k},$$

(1)

where $\xi$, $\eta$, and $\zeta$ are the vorticity components (see Holton 1992 for individual definitions) and $\mathbf{i}$, $\mathbf{j}$, and $\mathbf{k}$ are the unit vectors in the $x$, $y$, and $z$ directions, respectively. Standard manipulation of the horizontal momentum equations while neglecting contributions from friction, the Coriolis parameter, and the solenoidal term yields the following equation describing the time rate of change of vertical vorticity [$\partial \zeta / \partial t$ or the local generation (GENR)]:

$$\frac{\partial \zeta}{\partial t}_{\text{GENR}} = - \left( \frac{\partial \zeta}{\partial x} \frac{\partial u}{\partial x} + \frac{\partial \zeta}{\partial y} \frac{\partial v}{\partial y} \right)_{\text{HADV}} - \left( \frac{\partial \zeta}{\partial z} \frac{\partial w}{\partial z} \right)_{\text{VADV}} - \left( \xi \frac{\partial u}{\partial x} + \eta \frac{\partial v}{\partial y} \right)_{\text{CONV}} - \left( \frac{\partial w}{\partial x} \frac{\partial u}{\partial x} + \frac{\partial w}{\partial y} \frac{\partial u}{\partial y} + \frac{\partial w}{\partial z} \frac{\partial u}{\partial z} \right)_{\text{TILT}}.$$ 

(2)
where the first two terms on the right-hand side are the horizontal (HADV) and vertical (VADV) advection of vertical vorticity followed by the convergence (CONV, or stretching) and tilting (TILT) terms. Horizontal and vertical derivatives were computed from the Doppler-derived winds using center difference calculations (e.g., Pielke 2002). Terms involving vertical derivatives (j, h, VADV, and TILT) could not be computed at the lowest grid levels (below 1 km for cells A and B; below 1.5 km for cell C) because of limited Doppler observations.

1) HORIZONTAL VORTICITY

Current conceptual models of supercell evolution suggest that a primary mechanism for low to midlevel vertical vorticity production is the vertical tilting and stretching of horizontal vorticity associated with the environmental vertical shear (e.g., McCaul and Weisman 1996; Davies-Jones et al. 2001). Optimal production occurs when the low-level horizontal vorticity vector is largely aligned with the cell-relative wind vector (Davies-Jones 1984). In such cases, the streamwise horizontal vorticity is tilted into the vertical by an updraft, yielding a positive correlation between vertical velocity and vertical vorticity.

Figure 12 shows the low-level horizontal vorticity and cell-relative wind vectors in the vicinity of each miniature supercell updraft. Based on the environmental hodographs (Fig. 4), the shear vectors in the lowest 2 km were oriented toward the north-northeast, and thus the ambient horizontal vorticity vector in the low-level
inflow pointed west-northwest. As seen in Fig. 12, the vorticity vector orientations within 10 km of each updraft were consistent despite some local convective modulation. Moreover, the vorticity and storm-relative wind vectors immediately southeast of each updraft were largely aligned, indicating that streamwise vorticity dominated the inflow. Similar vector orientations were observed at all altitudes below 2 km (not shown). Horizontal vorticity magnitudes on the order of $5 \times 10^{-3}$ s$^{-1}$ were prevalent throughout the inflow layer and are consistent with the vertical vorticity magnitudes observed with each mesocyclone (Figs. 9–11). Therefore, in a manner consistent with the aforementioned conceptual models, the existence of low-level streamwise vorticity in the presence of preexisting midlevel mesocyclones supports the continued enhancement of vertical vorticity via tilting.

2) VERTICAL VORTICITY

Figure 13 shows the horizontal distribution of the vertical vorticity production terms [rhs of Eq. (2)] at 1.0, 1.5, and 2.0 km AGL in the vicinity of cell A’s mesocyclone. Below 2 km, the CONV term is positive, reaching a maximum value of $20 \times 10^{-6}$ s$^{-2}$ at 1.5 km within the mesocyclone. Above 2 km, the term becomes increasingly negative because of weak divergence. The TILT term at low levels is comparable in magnitude to the low-level CONV and TILT terms, occur in the 2–5-km layer. Finally, the HADV term exhibits a positive (negative) maximum on the west (east) flank of the mesocyclone at low levels consistent with simple advection by the environmental cell-relative winds. This couplet rotates 90° anticyclonically with height to a north–south orientation at 5 km AGL (not shown), which, in time, would produce a mesocyclone tilted to the northeast with height (as observed). Overall, low-level vorticity production within cell A was dominated by tilting and stretching, whereas midlevel vorticity production resulted from the vertical redistribution of cyclonic vorticity from lower levels.

The vertical vorticity production terms in the vicinity of cell B’s mesocyclone are shown in Fig. 14 at similar altitudes. As observed with cell A, the CONV term exhibits positive maxima below 2 km collocated with the primary mesocyclone (at $x = 340$ km). Maximum CONV exceeds $25 \times 10^{-6}$ s$^{-2}$ at 1.5 km AGL. Contributions above 2 km are negative. At low levels, the TILT term has positive contributions on the southeastern (upwind) flank but magnitudes are roughly half of those from the CONV term. Positive tilting contributions also diminish above 2 km. As with cell A, the VADV contributions within the mesocyclone are primarily negative below 2 km (i.e., below the vorticity maximum) but increasingly positive aloft with magnitudes comparable to the CONV term at low levels. The HADV term also exhibits the modest positive–negative couplet (with maxima on the mesocyclone flanks) that rotates anticyclonically with height in a manner consistent with the veering environmental winds. Interestingly, the shallow vorticity maximum (at $x \approx 346$ km), although likely a small-scale transient feature, exhibits qualitatively similar vorticity production structure as observed with the primary mesocyclone. Thus, low-level
vorticity production in cell B was dominated by stretching (with modest contributions from tilting), whereas midlevel production resulted primarily from vertical advection.

Despite complexity added by the more pronounced horizontal separation of the low- and midlevel mesocyclones, the vorticity production distributions in the vicinity of cell C (Fig. 15) are largely similar to those observed with the other cells. However, a few differences are notable. In particular, positive CONV maxima, collocated with the midlevel mesocyclone (at \( y \approx 45 \) km), extend up to 4 km. Positive tilting contributions also extend up to 5 km on the southeastern flank with a maximum of \( \sim 25 \times 10^{-6} \) s\(^{-2}\) at 2.5 km AGL. Such structure implies a deeper inflow layer for cell C, but midlevel outflow associated the small transient cell \( \sim 10 \) km to the southeast (see Fig. 7) also may have contributed.

Multiple Doppler analyses are not available to document supercell evolution. However, an indication of mesocyclone evolution can be deduced from the vorticity budget by examining the GENR term in (2) as a residual. Shown in Fig. 16 are the layer mean GENR terms for each cell (averaged through the altitudes shown in Figs. 13–15). Although the GENR maxima are not always collocated with their mesocyclone, GENR values of \( \sim 10–15 \times 10^{-6} \) s\(^{-2}\) overlap with each mesocyclone’s vertical vorticity maximum. Such local production rates would double the current mean vorticity values (5–8 \( \times 10^{-3} \) s\(^{-1}\)) within 10–14 min, implying mesocyclone and supercell intensification in the near future. As will be shown later, these supercells were indeed long lived with mesocyclones detected by the TLH radar as they approached the coast.

3) POTENTIAL BAROCLINIC CONTRIBUTIONS

Another mechanism believed responsible for the production of low-level vertical vorticity is the vertical tilting of baroclinically generated horizontal vorticity along cold outflow boundaries (Klemp and Rotunno 1983; Davies-Jones et al. 2001). The primary source region of such horizontal vorticity is typically downshear of the updraft, where low-level cell-relative inflow passes along the horizontal temperature (or \( \theta_c \)) gradient associated with the FFD outflow; significant baroclinic
Contributions along the RFD outflow have been noted (Adlerman et al. 1999) but often play a minimal role (e.g., Markowski et al. 2002). Numerical simulations suggest temperature gradients of $\sim 1^\circ C km^{-1}$ and maximum temperature deficits of $\sim 3–4^\circ C$ (or $\theta_e$ deficits of $\sim 8–12$ K) are necessary for significant baroclinic generation of horizontal vorticity (Klemp and Rotunno 1983; McCaul and Weisman 1996).

**Fig. 13.** Distributions of the HADV, VADV, CONV, and TILT terms in the vicinity of cell A at 1.0, 1.5, and 2.0 km AGL. Red (green) contours denote positive (negative) contributions to vertical vorticity production. Contours of $\pm 5, 10, 15, 20,$ and $25 \times 10^{-6}$ s$^{-2}$ are shown for each term. Also shown are the radar reflectivity (gray) shaded at the 15-, 20-, 25-, 30-, and 35-dBZ levels, the cyclonic vertical vorticity (blue) contoured at 2, 4, 6, 8, and $10 \times 10^{-6}$ s$^{-1}$, and the cell-relative wind vectors.
Although the three-dimensional distribution of such baroclinic contributions could not be determined from the available observations, an estimate of their magnitude can be inferred. Figure 17 shows the $w$, $T$, $T_d$, and $\theta_e$ observed at flight level (~2.5 km) during the rainband crossing. Note that the aircraft passed along the southern edge of any FFD but through the RFD region of cell B. Of the three prominent convective downdrafts ($w < -1.0$ m s$^{-1}$) encountered, two (at ~348 and ~366 km) contained subsaturated air ~1°–2°C warmer than their surroundings. The third downdraft (at ~353 km) contained only a modest cold anomaly.
(~1.5°C), some of which may be instrument wetting error, and a mean $\theta_e < 346$ K. Figure 18 shows the wind, $T$, $T_d$, and $\theta_e$ observed by the moored buoy (see Fig. 5). The distinct wind shift at ~2100 UTC marks the passage of the rainband axis over the buoy. Prior to this time, the buoy was located east of the band in the stratiform precipitation and was observing the band inflow as well as any cold FFD outflow. Minima in $T$ and $\theta_e$ were observed after 1800 UTC when any FFD outflow associated with cell C (or subsequent cells) would have passed over the buoy. However, maximum $T$ and $\theta_e$ deficits were only ~1°C and ~2 K, respectively, and

Fig. 15. As in Fig. 13, but in the vicinity of cell C at 1.5, 2.0, and 2.5 km AGL.
\( \theta_e \) remained >356 K. Thus, only limited evidence of any cold downdrafts or outflow boundaries was observed.

In agreement with McCaul and Weisman (1996), these observations suggest that cell-induced baroclinic contributions to miniature supercell evolution are minimal, even when dry midlevel air capable of supporting evaporatively driven downdrafts may be present. One plausible explanation is that the weaker updraft and weaker upper-level shear associated with miniature supercells result in less hydrometeor transport aloft and away from the cell and thus a smaller horizontal separation between the primary updraft and any FFD. The larger directional shear common at low levels may further limit any upwind area occupied by cold pools. Thus, in a hurricane, any cold FFD outflow might regularly be advected inside a rainband axis toward the storm center, limiting the necessary cold pools and thermodynamic gradients outside a rainband and upwind of any cells. Prior studies of hurricane rainbands and their convective cells (e.g., Powell 1990a,b; Cione et al. 2000) have observed near-surface cold pools but always inside the rainband axis. The cooler surface \( \theta_e \) observed by the GPS sonde (by \( \sim 5-7 \) K; see Table 2 and Fig. 18) is consistent and supports this explanation.

Contributions by preexisting baroclinic boundaries were more difficult to infer but may be significant. Assuming near-surface cold pools were prevalent inside and along the rainband axis (as discussed above), the rainband itself would embody an offshore baroclinic boundary with a shallow zone of enhanced vorticity and local solenoidal circulations. The numerical simulations of Atkins et al. (1999) suggest that any supercells propagating along such a boundary will develop stronger mesocyclones, with the boundary being an important ambient vorticity source.

6. Discussion

a. Subsequent offshore evolution and coastline tornadogenesis

The airborne dual-Doppler analysis at 1804 UTC clearly demonstrated that three distinct miniature supercells were embedded within Ivan’s outer rainband >100 km offshore. Figure 19 (combined with Fig. 2) provides a TLH-based perspective of each supercell’s subsequent evolution from their offshore locations at 1804 UTC until 2100 UTC, by which time each had moved onshore along the Florida Panhandle. During this period, the cells range from \( \sim 220 \) km (at 1804 UTC) to \( \sim 120 \) km (near 1845 UTC) from the radar, while the base-scan elevation decreased from \( \sim 5 \) to \( \sim 2 \) km as the cells approach the coast.

At 1804 UTC (Fig. 2c), none of the cells exhibited any distinct supercellular traits when viewed from TLH, because the base scan overshot any reflectivity signatures associated with the shallow mesocyclones. By 1831 UTC (Fig. 2d), cells A and B exhibited weak inflow notches along their southern flanks. At the same time, a new cell had appeared between the two, perhaps in response to cell B splitting or developing a transient outflow boundary. By 1852 UTC (Fig. 19), the new cell was merging with cell A, and cell C exhibited a distinct inflow notch. Over the next \( \sim 40 \) min, the radar imagery suggests that cell C underwent significant structural changes. At 1908 UTC, it no longer exhibited the notch and a new transient cell had developed along its northern flank. By 1929 UTC, the new cell was merging with cell A, and cell C exhibited a distinct inflow notch. Over the next \( \sim 40 \) min, the radar imagery suggests that cell C underwent significant structural changes. At 1908 UTC, it no longer exhibited the notch and a new transient cell had developed along its northern flank. By 1929 UTC, the new cell had propagated away and dissipated while the primary cell appeared reinvigorated and began to reacquire a weak inflow notch on its southern flank. Such behavior by cell C would imply the presence of at least modest downdrafts and an evolving outflow boundary. The behavior

![Fig. 16. Distributions of the layer mean GENR term in the vicinity of each cell. The domains and graph conventions are similar to Figs. 13–15.](image-url)
also was broadly consistent with a cyclic supercell (e.g., Adlerman et al. 1999), whereby the initial mesocyclonic updraft occludes (i.e., becomes cut off from its unstable inflow) and dissipates while a new mesocyclonic updraft develops just upwind (i.e., along the southern flank). Although strong downdrafts were not present in the Doppler wind fields; 1 h earlier, such downdrafts may have developed subsequently because of the ingestion of midlevel dry air (see Figs. 3, 7, 11, and 17). However, no direct observations are available to substantiate this hypothesis.

After ~1930 UTC, all three cells exhibited persistent inflow notches or hook echoes along their southern flanks as they paralleled and approached the coast from ~25 km offshore. During this period, prominent weak-echo vaults developed and midlevel rotational velocities rapidly increased (Baker et al. 2009), suggesting that the mesocyclonic updrafts intensified from interaction with the coastline. No radar-detected mesocyclones or tornadoes were observed with cell A as it moved onshore. However, a weak mesocyclone was detected briefly within cell B (at 1945 UTC) as it passed over a barrier island, and an F0 tornado was reported at 2040 UTC as the cell moved onshore south of Panama City Beach. Moreover, a moderate mesocyclone was detected continuously in cell C between 2022 and 2049 UTC while translating from ~10 km offshore to ~20 km onshore. The cell spawned at least three tornadoes between 2040 and 2055 UTC, the most damaging of which killed one person and injured seven with F1 damage in a commercial district of Panama City Beach.

b. Comparison to other tropical cyclones

Observations of offshore miniature supercells in tropical cyclones have been rare. Why did such supercells
develop offshore in this case? Was the environment exceptionally favorable? Compared to the mean offshore and onshore soundings from McCaul (1991) and Bogner et al. (2000), the CAPE values were typical (above average) for an offshore (onshore) environment, whereas the vertical shear and helicity were above average. The orientation and proximity of the rainband with respect to the upwind Florida Peninsula may have produced the greater shear and helicity through increased exposure to land (e.g., McCaul 1991). However, similar or greater helicity values have been observed in offshore tropical cyclones without the upwind exposure to land (Molinar and Vollaro 2008). Likewise, various forms of dry air intrusions often are observed well offshore (e.g., Dunion and Velden 2004; Curtis 2004). Therefore, although Ivan’s offshore environment was quite favorable, it does not appear to constitute a rare event.

Was the structure and organization of Ivan’s rainband significantly different? Compared to the gross structure of other offshore outer rainbands (e.g., Powell 1990a,b; Barnes et al. 1991), many characteristics of the precipitation and wind fields were similar. The most notable difference was in the organization of the band and convective cells. In particular, hurricane rainbands are often composed of multiple convective line segments spaced at 10–20-km intervals in the cross-band direction, whereas each line is composed of several distinct cells irregularly spaced ~10 km apart in the along-band direction. Ivan’s rainband, however, exhibited a single convective line with regular along-band cell spacing of ~20–30 km (see Fig. 2). It is unclear whether such organization permitted the development of multiple miniature supercells (by providing a continuous supply of warm moist air to each while

![Fig. 19. Base-scan radar reflectivity from TLH following the three miniature supercells as they approached the coastline and moved onshore between 1852 and 2044 UTC 15 Sep 2004. Each panel is 100 × 100 km and tic marks are shown every 25 km. The black circles denote radar-detected mesocyclones.](image-url)
limiting interactions with neighboring cells) or resulted from supercell dynamics. The numerical simulations of Bluestein and Weisman (2000) and James et al. (2005) provide support for the development hypothesis: cells initially spaced $>20$ km apart remained distinct and acquired supercell characteristics, whereas smaller cell spacings resulted in nearly continuous convective lines.

7. Summary and concluding remarks

A combination of airborne and land-based observations was used to document the structure and evolution of three miniature supercells in an outer rainband of Hurricane Ivan on 15 September 2004. In contrast to previous studies, the supercells were located more than 100 km offshore and beyond the Doppler range of coastal radars. Our specific findings include the following:

1) The offshore rainband environment was conducive to miniature supercell formation. Local SBCAPE ($>900$ J kg$^{-1}$), 0–6-km vertical shear magnitude ($>20$ m s$^{-1}$), and 0–3-km cell-relative helicity ($>200$ m$^2$ s$^{-2}$) equaled or exceeded mean values observed at similar radii in the right-front (or northeast) quadrant of hurricanes (McCaul 1991; Bogner et al. 2000), and convective inhibition was minimal.

2) Each offshore miniature supercell contained a mesocyclonic updraft that extended from the boundary layer up to $\sim$6–8 km AGL and was $\sim$5–7 km in diameter. Midlevel vertical vorticity (updraft) maxima exceeded $8 \times 10^{-3}$ s$^{-1}$. Such characteristics are consistent with onshore miniature supercells observed in association with tropical cyclone tornado outbreaks (e.g., Spratt et al. 1997; McCaul et al. 2004).

3) Updraft production and maintenance likely resulted from a combination of convergence, thermal instability, and shear-induced vertical perturbation pressure gradients.

4) At the time of the Doppler analysis, mesocyclone production resulted from the tilting and subsequent stretching of low-level horizontal streamwise vorticity into the vertical by the strong updrafts, as in midlatitude supercells (e.g., Davies-Jones et al. 2001). Evidence of baroclinic contributions from inflow along cell-generated outflow boundaries was minimal. Estimates of net vertical vorticity generation suggested that each mesocyclone was intensifying.

5) Each supercell persisted during the $\sim$3 h between their initial offshore examination and landfall. One cell showed signs of subsequently developing prominent downdrafts and undergoing cyclic mesocyclo-
genesis while still offshore. Two cells spawned reported tornadoes soon after moving onshore.

6) The rainband contained a single line of embedded convective cells regularly spaced at $\sim$20–30-km intervals. In contrast to typical offshore outer rainbands, such organization may have benefited supercell formation and maintenance by allowing individual cells to evolve in relative isolation from one another.

7) These observations build upon a growing body of evidence suggesting that miniature supercells often develop offshore in the outer rainbands of tropical cyclones.

The inability of current observational networks to regularly identify and monitor offshore supercell convection in tropical cyclones has limited our understanding of such events, as well as our ability to forecast any associated tornadogenesis. As a result, many open questions remain. In particular, how prevalent are supercells in the offshore outer rainbands? What characteristics of the tropical cyclone, rainband, and local environment support their formation? How does their structure and evolution differ offshore? How critical are preexisting boundaries (e.g., coastlines, dry air intrusions, and low-level baroclinic zones)? Ongoing work involves the examination of multiple outer rainbands embedded within a variety of offshore and nearshore tropical cyclone environments. Numerical simulations are also needed to isolate the relative roles played by various environmental features on supercell evolution.

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