Convective-Scale Downdrafts in the Principal Rainband of Hurricane Katrina (2005)

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ABSTRACT

Airborne Doppler radar data collected during the Hurricane Rainband and Intensity Change Experiment (RAINEX) document downdrafts in the principal rainband of Hurricane Katrina (2005). Inner-edge downdrafts (IEDs) originating at 6–8-km altitude created a sharp reflectivity gradient along the inner boundary of the rainband. Low-level downdrafts (LLDs) evidently driven by precipitation drag originated at 2–4 km within the heavy rain cells of each convective element. The IED and LLD were spatially separated but closely associated with the updrafts within the rainband. The IED was forced aloft by pressure perturbations formed in response to the adjacent buoyant updrafts. Once descending, the air attained negative buoyancy via evaporative cooling from the rainband precipitation. A convective-scale tangential wind maximum tended to occur in the radial inflow at lower levels in association with the IED, which enhanced the inward flux of angular momentum at lower levels. Convergence at the base of the downdrafts on the upwind end of the principal rainband contributed to the principal rainband growing in length. New updraft elements triggered by this convergence led to the formation of new IED and LLD pockets, which were subsequently advected downwind around the storm by the vortex winds while additional new cells continued to form on the upwind end of the band. These processes sustained the principal rainband and helped to make it effectively stationary relative to the storm center, thus maintaining its impact on the hurricane dynamics over an extended period.

1. Introduction

The principal rainband is a prominent feature of tropical cyclones (Willoughby et al. 1984). It consists of convective cells embedded in stratiform precipitation and spirals inward toward the storm center (Fig. 1a). It is larger than numerous other transient rainbands populating the storm, and is persistent and nearly stationary relative to the storm center. Despite its prominent appearance, the dynamic role of the principal rainband in the larger tropical cyclone remains uncertain. Barnes et al. (1983) and Powell (1990a) showed that the air spiraling inward toward the storm center is subjected to an overturning circulation when it encounters the principal rainband. Hence and Houze (2008) used high-resolution Doppler radar data to further document this overturning circulation (Fig. 1b). Their study emphasized the overturning updraft and how the upward motion builds and/or strengthens a midlevel jet that lies along the axis of the rainband. They and the previous investigators have also noted downdraft structures affiliated with the principal rainband. Air feeding a low-level downdraft (LLD) enters on the radially outward side of the principal rainband. The LLD transports low moist static energy air into the boundary layer. Another downdraft emanates from the upper levels on the radially inward side of the rainband. We call this feature the inner-edge downdraft (IED). The objective of the present study is to better document and understand these two types of downdraft structures and their possible roles in the dynamics of the hurricane. We use the same dataset and analyzed wind fields as Hence and Houze (2008): high-resolution measurements of air motion and precipitation structure obtained with the National Center for Atmospheric Research (NCAR) Electra Doppler Radar (ELDORA) and other data obtained during the 2005 Hurricane Rainband and Intensity Change Experiment (RAINEX; Houze et al. 2006, 2007). The specific objectives of this study are to

1) describe the characteristics of the LLD and the IED,
2) propose the forcing mechanisms that generate the two downdrafts, and
3) examine their immediate effects on the principal rainband and their probable impacts on the overall storm.

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Sections 2–4 will describe the data and methods of analysis. Sections 5–8 will identify the LLD and IED and describe the vertical velocity pattern in the principal rainband. Sections 9–12 will deduce the likely reasons for the LLD and IED, examine their immediate effects on the principal rainband, and suggest their possible impacts on the entire storm. Section 13 will draw conclusions about the role of the two downdrafts and suggest future directions for related research.

2. Doppler radar datasets

Under sponsorship of the National Science Foundation, the Naval Research Laboratory (NRL) P3 aircraft was deployed on 28 August 2005 as part of RAINEX to investigate the principal rainband of Hurricane Katrina. Airborne Doppler radar observations were taken from 2026–2145 UTC using ELDORA, which was on board the NRL aircraft. ELDORA is an X-band dual-Doppler radar (Hildebrand et al. 1996) that is noted for its high sampling resolution. The specific instrument settings for this mission allowed for an along-track data spacing of ~0.5 km within 40 km from the aircraft. ELDORA data were mapped onto a 0.5 km horizontal × 0.5 km vertical grid using a Cressman (1959) filter with a radius of influence of 0.75 km in the horizontal and 0.5 km in the vertical. Given a Nyquist wavelength of ~1 km, the influence radii chosen for the Cressman filter avoid errors resulting from the negative side lobes described by Trapp and Doswell (2000). The lowest grid altitude was 0.5 km. Vertical velocities were synthesized using several schemes. The first scheme was the “Gamache” variational method (described by Reasor et al. 2009). The remaining schemes used the NCAR methodology based on interpolation with REORDER (more information is available online at http://www.eol.ucar.edu/rsf/UserGuides/ELDORA/DataAnalysis/reorder/unixreorder.ps), and synthesis within the Cartesian Editing and Display of Radar Data under Interactive Control software (CEDRIC; Mohr et al. 1986).

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3. Principal rainband of Hurricane Katrina

Figure 2 shows the track of Hurricane Katrina, which began as a tropical depression over the central Bahamas (23°N, 75°W). Katrina made landfall in Miami, Florida (26°N, 80°W) as a category 1 (Saffir 2003) hurricane on
25 August 2005. It then tracked over the Gulf of Mexico for 3 days, where it intensified to a category 5 storm with a minimum central pressure of 902 hPa. Katrina was a very wide storm and made a second landfall east of New Orleans, Louisiana (29°N, 89°W) as a strong category 3 hurricane on 29 August 2005.

The shaded box in Fig. 2 shows the investigation area for the NRL P3 aircraft mission on 28 August 2005. At 2100 UTC, Katrina had a maximum sustained wind speed of 75 m s\(^{-1}\), a central pressure of 902 hPa, and was traveling to the northwest at 6 m s\(^{-1}\). Figure 3a shows the 2-km altitude reflectivity data from the ELDORA radar on the NRL P3 overlaid on the Geostationary Operational Environmental Satellite (GOES) visible data. During this time, Katrina contained a single eyewall (not shown in the radar data of Fig. 3) and a principal rainband on the eastern side of the storm. Just north of the eye, a secondary spiral rainband lay between the eye of the storm and the downwind end of the principal rainband. The volumes of radar data to the east and southeast of the eye show portions of the principal rainband. The current study focuses on the volume to the southeast, the most upwind volume of data. It is denoted by the red box in Fig. 3a and expanded in Fig. 3b. In this volume, a line of intense reflectivity was bounded by a sharp gradient of reflectivity along its inner edge.

4. Methods of analysis and general characteristics of the rainband

a. Convective–stratiform separation

Regions of convective precipitation consist of vigorous updrafts and heavy rain, seen on radar as locally vertically oriented cells of high reflectivity, whereas regions of stratiform precipitation are characterized by less vigorous vertical air motions and weaker rainfall (Houze 1997). The stratiform precipitation is less intense on average, is less variable in the horizontal, and is often signaled on radar by a bright band, where precipitating ice particles are melting (Houze 1993, chapter 6). To separate the convective and stratiform precipitation in the Katrina rainband, we use the convective/stratiform separation algorithm developed by Churchill and Houze (1984), Steiner et al. (1995), and Yuter and Houze (1997) (see appendix A).

Figure 4a presents the results of the separation applied to the data volume highlighted in Fig. 3. The separation appears to adequately capture the line of robust convective precipitation (in red) seen in Fig. 3b. For a statistical analysis of the classification, contoured frequency by altitude diagrams (CFADs; Yuter and Houze 1995b) were generated for the reflectivity and vertical velocity data in the convective and stratiform regions. CFADs display the height variation in data frequency.
distributions and provide insight into the precipitation and kinematic structure of clouds. Figures 5a,b show the reflectivity CFADs (in 5-dBZ bins) for the convective and stratiform regions, respectively. The stratiform reflectivity CFAD has the classic stratiform characteristics described by Yuter and Houze (1995b). It exhibits a bright band structure at the ~4.5-km level, and a very narrow distribution, indicating fairly uniform radar echo. The convective reflectivity CFAD, on the other hand, is distinctly different from the CFADs for convective regions of mesoscale convective systems. Convection in nonhurricane environments displays a broad distribution of reflectivity at all levels, with outliers of high reflectivity at the upper as well as the lower levels (Yuter and Houze 1995b). The convective echo CFAD in Fig. 5a does not have a broad echo distribution above the 8-km level; rather, it is almost identical to that of the stratiform echo in Fig. 5b at these upper levels. This

![Diagram](image1)

**Fig. 3.** (a) Composite plan view of ELDORA reflectivity data (dBZ) at 2-km altitude taken during 2026–2145 UTC 28 Aug 2005. Simultaneous GOES data are shown in the background. The red box indicates the data volume expanded in (b). (b) Expanded plan view of data volume indicated by the red box in (a), taken during the times 2026–2036 UTC.

![Diagram](image2)

**Fig. 4.** (a) Results of convective–stratiform separation algorithm applied to ELDORA reflectivity data shown in Fig. 3b. Red indicates convective regions, orange indicates stratiform regions, and green indicates weak-echo regions. Blue indicates regions of no reflectivity data. The overlying black lines are radii extending from the circulation center of Hurricane Katrina. The radii are labeled by the angle made with due east (0°). The white lines are cross sections of the principal rainband that are part of the total dataset for the current study. These radii are taken at angular intervals of 1.5° and represent 1/4 of the total cross-sectional dataset. (b) Plan view of vertical velocity data (m s⁻¹) at 4-km altitude. The red contour corresponds to the convective regions in Fig. 4a. Line A is a cross section taken at 42.2° south of due east. Line B is a cross section taken at 35.8° south of due east.
result reflects the fact that the convective cells in the principal rainband tend not to penetrate upward very far into the background stratiform echo at upper levels. The vertical extent of the convective cells is probably limited by the outflow from the eyewall. Below 8 km, the convective reflectivity CFAD differs markedly from the stratiform CFAD. The distribution has higher outlier values at all levels below 8 km, with peak reflectivities of 50 dBZ compared to 40 dBZ in the stratiform CFAD. The modal value of reflectivity below the 4.5 km (0°C) level was 38 dBZ in the convective echo as compared with 28 dBZ in the stratiform echo.

Figures 5c,d present CFADs of the vertical velocity data (in 1 m s^{-1} bins) in the different regions. Since the separation algorithm did not rely on the vertical velocity data, these CFADs provide an independent assessment of the vertical velocity field differences between the convective and stratiform echoes. Again, the differences in the distributions are most apparent at levels below 8 km, although the modal and outlier velocities in the convective echo are slightly but noticeably greater up to about 10 km, indicating that some convective cells penetrated to ~10-km height. Nonetheless, the primary differences between convective (Fig. 5c) and stratiform echoes (Fig. 5d) are seen at lower levels. There, the vertical velocity distribution is much broader in the convective echoes, which have outlier values reaching nearly 8 m s^{-1}, as compared with 3 m s^{-1} in the stratiform echo. The stratiform echo distribution is sharply peaked around zero. Thus, the convective regions throughout the lower and middle levels are seen to have a broad distribution of updrafts and downdrafts consistent with the robust overturning of air associated with active convection. On the other hand, stratiform regions tend to have weaker vertical velocities. This is reflected in Fig. 5b as the distribution is narrow at all levels.

The statistical analysis illustrated by the CFADs indicates that the convective–stratiform separation used in
this study adequately identifies the active convection of the principal rainband and distinguishes it from the background stratiform precipitation. In the remainder of the paper we seek a physical understanding of the processes associated with the convective echoes of the rainbands.

2. Radial cross-sectional analysis

The CFADs indicate only how echo structure varies in the vertical. To examine the horizontal variability of the principal rainband structure, we analyze the variation of the echo structure with distance from the storm center by means of a series of radial cross sections taken at azimuthal intervals of 0.375°. These radial cross sections cutting across the principal rainband capture the interaction of the hurricane’s secondary circulation with the convection in the principal rainband. The radial axis of each cross section extends from the storm center at 2031 UTC (the middle of the time frame of the radar data volume), which was determined to be 26.84°N, 88.94°W (based on a linear interpolation of the National Hurricane Center best-track data). Figure 4a shows a sample of the radii extending from the center of the storm through the rainband region, seen as black lines. The subsections of these radii selected for analysis of the convective line embedded in the rainband are shown as white lines. The portion selected for analysis was bounded by the outer edge of the convective zone and was at least 8 km in length across the convective zone. Since the active convection lay on the inner side of the rainband (Powell 1990a), an additional 8 km of data on the inner side of the rainband along each radius was added to each cross section selected for analysis. These additional radial segments include downdraft features that were related to the line but not exactly coincident with the convective reflectivity pattern. Cross sections selected for analysis were realigned to a common coordinate frame, where the origin of the coordinate system was assigned to the inner edge of each cross section. A total of 86 radial cross sections of the principal rainband were selected and analyzed in this composite coordinate system. Figure 4b is a plan view of the vertical velocity field at the 4-km level. The cross sections labeled A and B are used as examples in a later discussion.

c. Dropsonde analysis

We analyzed 18 dropsonde soundings taken on 28 August 2005 in the vicinity of Katrina’s principal rainband. In an analysis of the equivalent potential temperature \(\theta_e\) profile, soundings taken near the most active convection were consistent with Barnes et al. (1983), where the outer side of the rainband exhibited high \(\theta_e\) in lower levels and a minimum value of \(\theta_e\) in middle levels, and the inner side of the rainband was mostly uniform with height. Temperature and dewpoint profiles at the upwind end of the rainband exhibited a typical tropical sounding, with an unstable and moist boundary layer distinctly separate from the free atmosphere above it. The downwind end exhibited profiles typical of deep stratiform precipitation, nearly saturated and approximately moist adiabatic. In one case of two nearby soundings on the inner side of the principal rainband, one sounding taken in a small echo-free area had an “onion” skew T–logp pattern characteristic of downdrafts (Zipser 1977), while the other sounding was more uniform. The former sounding could have been taken in an inner-edge downdraft, but vertical velocity data were not available for confirmation. In general, the soundings were not inconsistent with any of the findings in this study or previous studies, but we did not glean any new insight into the rainband or downdraft structures.

5. Average updrafts and downdrafts

Data from the 86 cross sections were averaged to create a composite reflectivity and vertical velocity structure of the rainband. Although the convective cells are not a fixed distance from the classification boundary, the variation of this distance is minimal and is smoothed out in the averages. Averages of the updrafts and downdrafts were calculated separately and are shown in Figs. 6a,b. The average reflectivity values are displayed by the black contours. The outward-leaning reflectivity towers of the average convective cells in the principal rainband are evident in the average, similar to Fig. 1b. This average convective cell reaches values over 40 dBZ that rapidly drop off at the inner edge of the rainband. Figure 6a shows that the upward velocity of the overturning updraft (Fig. 1b) encompasses a broad area on the inner side of the rainband, extending from near the surface to 10-km altitude and spanning 10 km in horizontal distance. The updraft reaches a maximum average speed of over 4 m s\(^{-1}\) between 3- and 5-km altitude.

Figure 6b reveals several pockets of downdrafts. One mean downdraft is located radially outward of the updraft core in the middle of the reflectivity tower (15 km from origin) and has a maximum of 2.4 m s\(^{-1}\) at 3-km altitude. This is the LLD described by Hence and Houze (2008) and shown within the reflectivity tower of Fig. 1b. It extends from near the surface up to 4-km altitude and spans 5 km in horizontal distance. The most prominent downdraft, which is separate from the LLD, is located radially inward of the updraft core and centered along the sharp reflectivity gradient defining the inner edge of the principal rainband. It extends from near the
surface up to 9-km altitude and spans 8 km in horizontal distance. This feature is the IED depicted in Fig. 1b. The IED has two average maxima: one reaching 2.5 m s$^{-1}$ between 4- and 5-km altitudes and another reaching 3 m s$^{-1}$ between 1- and 2-km altitudes. A small maximum occurs at 3 km altitude, but it appears to be associated with the low-level maximum of the IED. Near the origin, a downdraft maximum occurs at 2.5-km altitude. This downdraft is likely associated with the weak convective cells that occur 10 km inside the downwind end of the convective line seen in Fig. 3b. The last downdraft maximum occurring at 26 km from the origin is likely

![Fig. 6](image-url)
due to retrieval error since the aircraft flew along the outer edge of this reflectivity band.

6. Frequency distribution of vertical velocity

The frequency of occurrence of updrafts and downdrafts is examined. The black contours in Fig. 7a are the percentage frequency of occurrence in the cross sections where the updrafts were $\geq 3.0$ m s$^{-1}$ (these updrafts will be referred to as strong). The frequency contours are overlaid on the average reflectivity. The most frequent strong updrafts, contained within 32.5% of the cross sections, occurred in roughly the same location as the updraft maximum. The black contours in Fig. 7b show the percentage of occurrence of downdrafts $\geq 1.5$ m s$^{-1}$ (these downdrafts will be referred to as strong). A lower threshold is used since the average downdraft was weaker than the average updraft. According to Fig. 7b, the strong downdrafts also occurred less frequently than the strong updrafts. The distinction between the LLD and IED can again be clearly seen, as they are both confined to their respective locations seen in the average. Strong LLDs reached a frequency of 15% and strong IEDs reached a frequency of 20%.

The frequency distribution of IEDs may be analyzed further using a CFAD of the mass-transport-weighted vertical velocity (Yuter and Houze 1995c). This CFAD shows which downdraft velocities contribute most to the overall downward mass transport. The air density used for the mass transport calculations is from an upper-air sounding at Tallahassee, Florida, at 1800 UTC 28 August 2005. Figure 8 shows the mass-transport-weighted vertical velocity CFAD of all downdrafts (in 1 m s$^{-1}$ bins) located between 8.5 and 12.5 km from the origin, which corresponds to the IED region. Three distinct maxima are evident within the IED. At 4.5-km altitude, the 3 m s$^{-1}$ velocities associated with the upper portions of IEDs yield a maximum in vertical mass transport. At 1 km, separate maxima in vertical mass transport occur at 2 and 4 m s$^{-1}$, associated with the lower portions of IEDs. The mass transport CFAD is consistent with Fig. 6b. The minima in vertical mass transport that separate each of these three downdrafts underscore the uniqueness of each downdraft and suggest that their underlying processes are physically distinct.

How do the frequency distributions of the midlevel and low-level IED downdraft speeds relate to each other? By conservation of mass, air sinking in the middle levels has a direct impact on the air column below. A conditional probability analysis is applied to the same IED data to reveal the connectedness in time between the two levels. Figures 9a,b present the probability distribution of 1-km level maximum IED speeds when the 4.5-km level IED reaches a certain maximum speed. Figure 9a shows that when the maximum 4.5-km IED is $\leq 3$ m s$^{-1}$, only 19% of the maximum 1-km IEDs reach speeds $> 2$ m s$^{-1}$. Figure 9b shows that when the 4.5-km IED is $> 3$ m s$^{-1}$, a much larger portion, 60%, of the 1-km IEDs reach speeds $> 2$ m s$^{-1}$. This shift in distribution is found when comparing any midlevel altitude to any low-level altitude. In summary, a weak midlevel IED is most likely accompanied by a weak (1–2 m s$^{-1}$) low-level IED or none (0 m s$^{-1}$) at all. Once the midlevel IED exceeds a certain threshold (3 m s$^{-1}$ in this case), the probability of an accompanying significant low-level IED becomes much higher.

7. Temperature and pressure perturbation fields

Given the reflectivity and three-dimensional wind field, we were able to perform a steady-state thermodynamic retrieval, as described in appendix B, on a portion of the data volume. Using the equations of motion, thermodynamic equation, and water continuity equations, the retrieval yields fields of buoyancy and pressure perturbations relative to an environmental sounding (the Tallahassee sounding used in section 6). Since solutions are obtained by minimizing errors over the whole spatial domain, we narrowed the domain to be all data west of 87.49°W in order to avoid the retrieved wind errors along the path of the aircraft. We present the results in cross sections oriented in the radial, cross-band direction. In each cross section, the retrieved buoyancy and pressure perturbations were averaged for each altitude, and the averages were then subtracted from the individual retrieved buoyancy and pressure perturbation at each data point. This was done to best capture local perturbations relative to the profile of base state pressure in the vicinity of the region of radar retrieval analysis rather than the far-field sounding. The perturbations seen near the center of the cross section are then due to the convection in the rainband and not the hurricane environment. Since the abbreviated data volume cuts out a portion of the downwind end of the rainband, only the upwind half of the cross sections (between 54.9° and 35.0° south of due east) are considered in the analysis.

Figure 10a shows the average field of temperature perturbations in the cross sections along with the average downdrafts as black contours. A large swath of positive temperature perturbations extends from 2- to 7-km altitude and spans the width of the mean downdrafts and updrafts (Fig. 6a). Since positive temperature perturbations indicate buoyant air, the dominance of rising air in the average is consistent with the relative strength and frequencies of updrafts compared to downdrafts. In the lowest 2 km, the air becomes cooler as it travels radially...
inward from outside the rainband, which is consistent with previous studies (Barnes et al. 1983; Powell 1990b). Figure 10b shows the average field of pressure perturbations in the cross sections. Because the pressure perturbation field is the deviation from a horizontally uniform basic state, the pattern in Fig. 10b shows a general decrease in perturbation pressure traveling radially inward, which reflects the mean vortex contribution to the pressure perturbation field. This general decrease in pressure with decreasing radius associated with the mean vortex does not affect our interpretations because we are interested only in the vertical gradient of the pressure.
boundary conditions, $p_D$ was set to zero at the lowest and highest altitudes. Neumann boundary conditions were applied to the $x$ and $y$ boundaries, and their values were calculated by inverting the divergence operator on both sides of (2). The resulting field of $p_D$ appeared not to be sensitive to variations in the interpolation scheme for missing data. In a separate test calculation, we solved (2) by assuming $\nabla^2 p_D = -p_D$ and using scale lengths appropriate to the updraft and downdraft cores in the dataset. The resulting values of $p_D$ were similar to those calculated by the numerical Laplacian solver. Given $p'$ and $p_D$, the buoyancy pressure perturbation field was solved via (1). In another separate test calculation, buoyancy pressure perturbations were computed for a buoyant box of $+1$ K, similar to the retrieved buoyancy values from Fig. 10a. The resulting values for $p_B$ were similar to those $p_B$ values in the location of the downdrafts as calculated from (1).

Figures 11a,b present the average cross-sectional fields for $p_B$ and $p_D$ respectively. Note that the $p_B$ field still contains the mean vortex contribution to the pressure perturbation; however, these values are negligible near the center of the cross section. Furthermore, it is the vertical gradient of these perturbations, not the perturbations themselves, that is most important to forcing of the air. The vertical pressure gradient accelerations of the $p_B$ and $p_D$ fields are shown in Figs. 11c,d. In the location of the downdrafts (8–16 km from the origin), the $p_B$ field resembles the $p'$ field (Fig. 10b), as the average values of $p_D$ are smaller in comparison. Between 3- and 6-km altitude at the IED location, $p_B$ ranges from +20 to −20 Pa, creating downward acceleration. Values of $p_D$ reach an average of $\sim +10$ Pa in the same area creating a faint signal of downward acceleration between 2- and 3-km altitude, but not nearly as robust as the downward accelerations associated with $p_B$. In individual cross sections, accelerations owing to $p_D$ reach maximum values of 0.03 m s$^{-2}$, comparable to the average for buoyancy accelerations; however, this only occurs for one of the downdraft cores. A buoyancy-induced pressure minimum occurs in the lower levels (0.5–2 km) at $\sim 15$ km from the origin (Fig. 11a), resulting in downward acceleration near the center of the LLD (Fig. 11c). Powell (1990a) also noted a pressure minimum in the same relative location, and suggested buoyant and dynamic forcing as the two possible causes. Our calculations indicate that buoyancy is the dominant factor producing the minimum in this location.

8. **Horizontal distribution of vertical velocity**

In Fig. 4b, the downdraft cores were more likely to be IEDs rather than LLDs, since the average LLD occurred
below the 4-km level (Fig. 6b). The principal rainband consisted primarily of an intermittent pattern of updraft and downdraft cores. These cores were 5–10 km in diameter, affirming that the updrafts and IEDs were both convective-scale phenomena (i.e., the IEDs were related locally to the dynamics of convective cells rather than to the broader dynamics of the hurricane vortex). The updrafts were slightly larger and stronger than the downdrafts, just as seen in the composite cross sections (Fig. 6). A miniature version of these alternating cores is also present in the developing outer convective line of the principal rainband seen near the center of the data volume.

We investigate the pattern of updraft and downdraft cores by focusing again on the upwind half of the data volume, where the updraft and downdraft cores were newer and their alternation was generally more robust. Figure 12a presents the average downdrafts and reflectivity of the upwind half of the cross sections. When compared to Fig. 6b, both averages show a clear reflectivity tower, a sharp reflectivity gradient, an LLD, and upper and lower portions of the IED. Thus, this sample of the dataset is an adequate representation of the total data volume.

An autocorrelation of vertical velocity was performed on these cross sections in the along-band direction. Each data pixel corresponding to a certain distance from the origin and altitude was correlated with the same pixel from different cross sections over a range of cross-sectional lags. Figure 12b presents the autocorrelation coefficients for a lag of four cross sections. A lag of four roughly corresponds to a horizontal distance of 4.5 km, which is about the same distance between updraft and downdraft cores. The statistical significance of the autocorrelation was determined using the two-sided Student’s t statistic:

\[ t = \frac{r \sqrt{N - 2}}{\sqrt{1 - r^2}} \]  

where \( r \) is the correlation coefficient and \( N \) is the number of independent samples. Here \( N \) was determined by the formula of Bretherton et al. (1999):

\[ \frac{N^*}{N} = \frac{(1 - r^2)}{(1 + r^2)} \]  

where \( N^* \) is the number of cross sections. The null hypothesis of this significance test was that the true autocorrelation coefficient is zero. The black, dashed contour marks the areas where the correlations are statistically significant at the 95% confidence level. We focus on the large area of statistically significant negative correlations (ranging from −0.3 to −0.7) along the reflectivity gradient, extending from near the surface to about 6-km altitude. This area largely corresponds to the area covered by the IED when compared to Fig. 12a.

The high negative autocorrelation means that for every IED, an overturning updraft likely exists in the same region relative to the reflectivity tower 4.5 km upwind or downwind. This finding strongly suggests a physical relationship between the IEDs and updrafts. Although it is less clear, the region of negative correlations at 20 km from the origin and 7-km altitude may also be a result of alternating IEDs and updrafts. The individual cross sections show that weak downward motion associated with the midlevel IED sometimes occurs in this region.
region, and Fig. 6a shows that the average updraft also extends into this region.

9. Forcing of the downdrafts

Downdrafts in ordinary convective storms have four basic, well-documented forcing mechanisms. Zipser (1977) observed and classified two distinct types of downdrafts that occur in squall-line systems. Within the heavy precipitation zone of the squall line, intense, convective-scale saturated downdrafts are initially forced by precipitation drag, and the air becomes negatively buoyant and accelerates downward because of continuous evaporative cooling. Beneath the precipitating anvil cloud, a
much broader, mesoscale downdraft forms due to cooling largely from evaporation. Other studies document a third type of downdraft that occurs in the upper levels of deep convective clouds as a gravity wave response to the penetrative updraft tower (Palmén and Newton 1969; Biggerstaff and Houze 1993; Yuter and Houze 1995a,b).

The buoyant air within the storm creates buoyancy pressure perturbations ($p_B$, Pa) and forms a buoyancy pressure gradient acceleration (BPGA) field. Figure 13 illustrates the BPGA field that a rising air parcel exerts on its environment, causing air to be forced laterally away from the top of the parcel, down the sides of the parcel, and back toward the bottom of the parcel. The BPGA field within the parcel itself is downward, acting against its upward motion. A fourth type of downdraft forms from dynamically induced pressure perturbations ($p_D$, Pa) (section 7). Rotunno and Klemp (1982) showed that directional shear in a mature thunderstorm creates dynamic pressure perturbations around the main updraft core. As a result, a dynamic pressure gradient forms and sinking air counteracts the updraft on one side of the storm. This last type of downward pressure gradient force occurs most notably in supercell thunderstorms, which have especially strong rotationally induced minima of $p_D$.

Figure 14a presents cross section A from Fig. 4b. A pronounced LLD reaches speeds of 3 m s$^{-1}$ and is embedded completely in a 35-dBZ reflectivity tower. Since the downdraft is located in the lower levels and in heavy precipitation, it is most likely forced by precipitation drag, in the manner of the convective-scale-saturated downdrafts described by Zipser (1977).

Figure 15a shows cross section B from Fig. 4b with reflectivity values in color and vectors of the wind in the plane of the cross section. The corresponding vertical velocity values are plotted in Fig. 15b. The IED circulation flanks a robust reflectivity tower, while the overturning updraft is located within the inner-edge...
portion of the reflectivity tower. The IED originates between 6- and 8-km altitude, where the initial forcing that causes the IED must exist. The IED cannot be initially forced by evaporative or sublimational cooling because such cooling would induce a downdraft with negative buoyancy. Figure 10a shows positive buoyancy in the upper portion of the IEDs. Moreover, ice crystals flow out from the eyewall in every direction; any
The downdrafts induced by cooling from evaporation or sublimation in the mid- to upper levels would appear everywhere, rather than concentrating at the locations of the IED along the rainband. The IED cannot be initially forced by precipitation drag because heavy precipitation is not falling at this altitude on the inner side of the rainband. Last, dynamic pressure perturbations are an unlikely cause for the initial forcing since, even despite the weak magnitudes, the strongest dynamic pressure gradient within the column of the IED occurs at the 3-km level, which is much lower than the downdraft’s origin.

The remaining explanation, which is also best supported by the observations, is the buoyancy-induced pressure gradient acceleration. As observed by Biggerstaff and Houze (1993) and Yuter and Houze (1995b), the updraft cores in the rainband exert a BPGA field of the form illustrated in Fig. 13, but in three dimensions. Since the strength of the BPGA field is dependent on the strength of the buoyancy of the updraft, the field is strongest around the updraft core on the sides where the horizontal gradient of vertical velocity is the strongest. A closer look at Fig. 4b shows that downdrafts tend to be located along the sides of updraft cores where the buoyancy gradient is likely strongest. In Fig. 15b, the positive vertical velocity increases from 0 to ≥10 m s⁻¹ over a horizontal distance of less than 2 km. A strong downdraft core is adjacent to this strong gradient, and the two cores extend roughly to the same altitude. In the thermodynamic retrieval, downward acceleration from the average buoyancy pressure perturbations overlays the IED between the 4- and 6-km altitude, which is close to the origination level. Compared to the averaged cross section, individual cross sections of the retrieval results appeared very noisy, which could be attributed to the time disconnect between forcing mechanisms and resulting motion.

If the BPGA field explains the initiation of the IED, how does the IED extend into the lower levels? As described by Sun et al. (1993), an air parcel mechanically forced down by the BPGA field becomes positively buoyant, causing it to return to its original altitude. The air below the parcel in the same column is pushed down also, but at weaker speeds than the parcel itself. This scenario is hypothesized to be the case for the weak midlevel IEDs from Fig. 9a, where little or no low-level IED exists for the majority of weak midlevel IEDs. However, the strong low-level IEDs observed in Fig. 9b must be explained by another factor. When compared with the buoyancy accelerations and the resulting midlevel downdraft speeds, the dynamic accelerations do not appear strong enough to solely create the observed low-level downdraft speeds. In section 10, we show that evaporation is occurring along the edge of the rainband, which leads us to believe that the dominant factor causing the low-level IEDs is evaporative cooling from the rainband precipitation.

The points above combined with the IED CFAD and conditional probabilities from section 6 provide evidence for the following extension of the IED-forcing hypothesis: as the midlevel IED becomes stronger, more air below it is forced downward. With some further forcing by dynamic pressure perturbations, the sinking air juts into the adjacent heavy precipitation of the rainband and evaporates the precipitation. Eventually, at some level, the air cools enough for it to become negatively buoyant and accelerate downward. The BPGA field forcing and subsequent evaporative cooling thus explain the midlevel and the low-level IED maxima observed in the composite cross-sectional field.

10. Effects of the IED
a. Sharp inner-edge reflectivity gradient

The sharp inner-edge reflectivity gradient of Katrina’s principal rainband is seen in Fig. 3b. From Fig. 6b, the cross-sectional average shows that the IED occurs right along this gradient. At about 7 km, the 20-dBZ contour drops steeply where the IED is strongest, marking the upper extent of the reflectivity gradient. The gradient is strongest near the surface, where average reflectivities (rain rates) transition from >40 dBZ (~20 mm h⁻¹) to <30 dBZ (~3 mm h⁻¹) over a 3-km distance. Previous studies of other hurricane rainbands identify similar inner edges with a clear lane on the inner side of the rainband (Willoughby et al. 1984; Powell 1990a). Barnes et al. (1983) discussed its temporal consistency in Hurricane Floyd (1981), while noting that there were short-lived periods when the inner edge of the rainband was ill defined.
Hence and Houze (2008) suggested that the IED is the cause for the sharp reflectivity gradient on the inner edge of the rainband because of their concurrent locations. A closer look at Fig. 15a shows evidence to support this hypothesis. In the cross section, a weak bright band (∼25 dBZ) at 4.5 km is located adjacent to the intense reflectivity tower (at ∼10 km on the horizontal axis) as a result of melting ice crystals originating from the eyewall or the rainband itself. Below it, the reflectivity values gradually decrease toward the surface, reaching ∼15 dBZ. This suggests that precipitation is evaporating as it falls, which sharpens the reflectivity gradient defining the edge of the rainband. The evaporation is likely a result of subsidence and drying in the IED. Viewing other cross sections verifies that IEDs are repeatedly seen over dry areas such as in Fig. 15a.

FIG. 14. Cross section A from Fig. 4b, located 42.2° south of due east. The horizontal axis is the same as in Fig. 6. (a) Vertical velocity (m s⁻¹) with reflectivity values overlaid as black contours (dBZ). (b) Divergence (10⁻³ s⁻¹) overlaid by dual-Doppler wind vectors in the cross-section plane.
Although the IEDs are convective-scale cellular structures in the plan view, the sharp reflectivity gradient is consistent along the rainband. Thus, one must ask the question, why is the sharp reflectivity gradient located continuously along the entire band rather than just where the IED cores are located? In a hurricane environment, the subsiding, drier air is advected along the rainband faster than a single-core updraft can form precipitation. Given the tangential wind speed and distance between draft cores, it would take just over 2 min for dry air to be advected from a downdraft core to an updraft core. In comparison, it would take at least 13 min for precipitation particles within an updraft core to grow and fall out in order to relax the surface reflectivity gradient.

b. Low-level wind maximum

Figure 15c presents the field of tangential winds in a rainband cross section. The secondary horizontal wind maximum (SHWM) described by Hence and Houze (2008) is located between 3- and 7-km altitude and radially outward from the overturning updraft. They suggest that this SHWM is a kinematic consequence of the vertical wind shear and the updraft. On the inner side of the IED is another local tangential wind maximum between 2- and 3-km altitude, where wind speeds
reach 44 m s\(^{-1}\). This low-level wind maximum (LLWM) is repeatedly seen along the rainband and most often occurs, or is most pronounced, where it coincides with the occurrence of an IED. We find that 73% of the cross sections with an LLWM also exhibit an IED, which suggests an association between the LLWM and IED. Figure 16 is a composite tangential wind cross section that Barnes et al. (1983) derived from data taken in and around the principal rainband of Hurricane Floyd (1981). This cross section displays an LLWM at the 1.5-km level on the inner side of the rainband, similar to that seen in Hurricane Katrina. Barnes et al. (1983) also inferred the existence of a persistent downdraft in mid-to upper levels on the inner edge of the rainband, which could possibly have been the IED. However, they did not have the detailed data necessary to determine that the upper-level subsidence in their case was convective scale and correlated with convective updrafts.

The fields of vertical vorticity and divergence lead us to hypothesize a dynamical relationship among the IED, overturning updraft, and LLWM, which would serve as an extension to the kinematic model of the principal rainband presented by Hence and Houze (2008). As depicted in Fig. 1b, tangential wind speed (\(u\)) increases with height (\(\partial u/\partial z > 0\)) at the lower levels where air is flowing into the rainband. This wind shear is characteristic of the air flowing into the rainband and the basic-state flow at the location of the rainband. The radial inflow thus contains substantial horizontal vorticity \(\xi\) that is then tilted up within the rainband updraft, producing a rotating updraft with positive vertical vorticity. Simultaneously, the rotating updraft tilts the basic static shear and forms a vertical vorticity couplet along the inner edge of the rainband, as proposed by Powell (1990a). These anomalies would combine to create a total vorticity pattern where the positive vorticity center of the vorticity couplet remains dominant within the updraft but with a negative vorticity anomaly on the inner edge of the rainband. Figures 15d,e show the vertical vorticity \(\zeta\) and divergence fields for the example rainband cross section. The vertical vorticity couplet (located 10–15 km from the origin and below 2-km altitude) is likely a result of the aforementioned tilting processes, while the positive vorticity core is aligned with the updraft (Fig. 15b). According to the Hence and Houze (2008) model (Fig. 1b), this positive vorticity is stretched and intensified by the overlying region of convergence (Fig. 15c), and is also advected upward by the updraft. At the same time, the negative vorticity core at 10 km from the origin is collocated with the IED. This suggests that the IED advects the negative vorticity downward, causing the vertical flux of negative vorticity to converge in the lower levels. Calculation of the terms contributing to the instantaneous local derivative of vorticity shows that the tilting and vertical flux convergence terms are of the same order of magnitude, and occur in the same location as the stronger and larger negative vorticity core seen in Fig. 15d at 11 km from the origin. The vorticity features seen in Fig. 15d are also present in other cross sections along the rainband. The negative vorticity maximum is thus manifested as decreased tangential velocity on its radially outward side and increased tangential velocity on its radially inward side, which is consistent with Fig. 15c and results in the LLWM that is associated with the IED. The negative vorticity located radially outside the updraft (~16–21 km from the origin) appears to be associated with the outer edge of the SHWM, which anomalously extends down to the surface in this cross section.

In the cases in which only an IED exists with no updraft, there is no sign of an LLWM, which further indicates that tilting from the updraft is the main source of the negative vertical vorticity associated with the LLWM. Moreover, in the cases where the LLWM exists but the IED does not, the overturning updraft is exceptionally strong. These intense updrafts also create intense vertical vorticity couplets, such that the negative vorticity core is strong enough generate a LLWM without the vertical flux convergence from an IED.

Throughout the dataset, the updrafts and IEDs associated with the principal rainband appear to form a dynamical boundary between the strong low-level winds, inward of the principal rainband, and the midlevel SHWM, located within the band itself. The strong low-level winds are an integral part of the mesoscale vortex circulation of the hurricane. The SHWM, which is a vertical wind maximum in the 4–6-km altitude range, is
inconsistent with hurricane vortex dynamics, which specify that the maximum wind speeds occur in the lower levels. The disparity of tangential wind vertical distribution on either side of the updraft and IED supports the hypothesis of Willoughby et al. (1984), who described the principal rainband as a boundary between the hurricane vortex and the surrounding environment. That the hurricane vortex cannot be responsible for the SHWM within the principal rainband, suggests that the region radially outward from the principal rainband is more susceptible to convective-scale and environmental influences.

11. Possible impacts of the downdrafts

a. Increasing the angular momentum of the hurricane vortex

The IEDs associated with the principal rainband may constitute a positive feedback to the overall storm. As seen in the example of Fig. 15c, the LLWM is located in the low-level radial inflow. The increased angular momentum from the LLWM must be advected inward. The local anomalous increases in low-level winds, which are aided by IEDs, thus act to strengthen the overall storm by contributing to an increase of the flux of angular momentum toward the storm center, where it is manifested as increased tangential winds.

b. Growth and sustenance of the principal rainband

Rainband formation in hurricanes has traditionally been explained in terms of processes occurring on the vortex- and mesoscales. On the vortex scale, some studies suggest that rainbands result from the asymmetric distribution of low-level convergence due to the translating storm vortex interacting with its environment (Shapiro 1983; Willoughby et al. 1984). On the mesoscale, other studies suggest that rainbands are the manifestation of waves (inertia–gravity waves or vortex–Rossby waves) emanating from the storm center and propagating outward (e.g., Kurihara 1976; Montgomery and Kallenbach 1997). In the case of a principal rainband of the type studied here, convective-scale processes may also play a role in its growth and/or sustenance. Several studies have determined that the upwind portion of rainbands consists of active convection, while the downwind portion consists of decaying convective cells and stratiform precipitation (e.g., Atlas et al. 1963; Barnes et al. 1983; Hence and Houze 2008). As observed by Parrish et al. (1984), this signifies the upwind development of cells and upwind growth of the entire rainband. Powell (1990a) suggested that the minimum in pressure perturbation observed radially outward of the updrafts may also play a role in this upwind growth. This study neither proves nor disproves the theories involving larger-scale processes; rather, the new observations presented here relate to a simple convective-scale hypothesis that may also play a role in the growth and sustenance of the principal rainband.

Several observational studies hypothesize that spreading downdrafts help force updrafts within the rainbands of a hurricane (Ishihara et al. 1986; Barnes et al. 1991; May 1996). Figure 14b illustrates the divergence field of the LLD in a cross section corresponding to Fig. 14a. A region of divergence lies in the lowest levels below the LLD (at radial distance 18 km), and in the lowest levels below the IED (Fig. 15e, at a radial distance 10 km). In the region of the hurricane upwind of the principal rainband, the vortex circulation dominates the airflow with little environmental influence. In the boundary layer, turbulent mixing raises the moist static energy of the air as it interacts with the warm ocean below. When this rapidly moving air encounters near-surface negatively buoyant downdraft air spreading out away from a convective precipitation cell, convergence occurs in the lower levels and convective triggering becomes favored on the upwind side of the diverging downdraft. New updraft cores would lead to convective cells developing on the upwind end of the rainband.

Figure 17 shows the evolution of Ophelia (2005), as seen by the operational radar at Melbourne, Florida, during its early tropical storm phase. At 1700 UTC, Ophelia had a well-defined center, stratiform precipitation to the west and north, and discrete convective cells to the east, showing signs of a nascent spiraling principal rainband. At 1805 UTC, the cells to the east remained in roughly the same location, but were slightly larger than before. At 1912 UTC, the cells start to connect to form a single spiral rainband, showing a significant growth of convective cells upwind. At 2017 UTC, the principal rainband is a well-defined spiral band of robust convection that has grown even farther upwind, reaching the region southwest of the storm center. Thus, in just a few hours, convective cells grew systematically upwind in areas that were previously clear, forming a single rainband. The way in which Ophelia’s rainband grew suggests that the convective cells themselves may have altered the rainband’s environment such that growth of new cells could occur.

Figure 18 illustrates a possible scenario showing the interaction between updraft and downdraft cores at the 2-km level of a principal rainband. At this level, downdraft cores consist of both IEDs and LLDs. The principal rainband, outlined by the 25-dBZ contour, consists of intermittent updrafts and downdrafts, as seen in the observations. At the tail end of a preexisting downdraft core, a region of convergence forms on its upwind side where the low-level vortex flow encounters the denser
air associated with the downdraft, thus providing the upward forcing necessary to create a new buoyant updraft core. At the same time, this downdraft core may be acting to strengthen the updraft core immediately downwind. The lowest 3 km of the principal rainband has positive vertical wind shear where the midlevel jet exists. This circumstance suggests that upward motion is enhanced on the downwind side of the downdraft (Rotunno et al. 1988). This would allow for continued growth of updraft cores as they are advected immediately downwind, as illustrated by the stronger updrafts downwind in Fig. 18. In the far downwind end of the rainband, the older convective cells would still collapse into stratiform precipitation as newer cells upwind block the inflow of warm, moist air along the rainband.

The counteracting effects of upwind development and downwind advection constitute the discrete propagation of convective cells, similar to squall-line behavior studied by Newton and Fankhauser (1964). In the absence of outside influences, the result would be a principal rainband that remains sustained and quasi-stationary relative to the storm center for extended periods of time, which is consistent with observations in previous studies. This has led to the concept that the principal rainband tends to be roughly stationary relative to the storm (Willoughby et al. 1984). Outward radial propagation of the principal rainband is also sometimes observed, a behavior that may be better explained by wave dynamics. Nonetheless, the hypotheses put forward here can help explain the sustenance and continued quasi-stationary position of the principal rainband through upwind growth.

12. Conceptual model of convection within the principal rainband

Figure 19 summarizes in a conceptual model the key aspects of the LLD and the IED and their roles in the principal rainband. This cross section cuts across a convective cell within the principal rainband. The cell is characterized by a radially outward leaning tower of
reflectivity \(>\sim 40\ \text{dBZ}\). For a given radial cross section, the airflow structure associated with this cell can contain one or any part of the broad arrows shown in the illustration. The IED is shown in the conceptual model as the series of darker gray, solid arrows located along the inner edge of the principal rainband, where each arrow represents a different process at that stage in the flow pattern. The IED begins around the 8-km level in the layer of radial outflow from the vortex circulation. As the airflow approaches the inner side of the rainband, it is forced downward by the BPGA field created by a buoyant overturning updraft located within the principal rainband. This initial forcing can cause downward motion to occur upwind of the updraft, downwind of the updraft, or radially adjacent to the updraft on its inner side. The next step in the IED development is determined by the strength of the BPGA field, which dictates how much and how strongly air is pushed down into the heavily precipitating rainband. If the air pushed downward enters and remains in the precipitation region (marked by the radar reflectivity) long enough, it cools from evaporation, becomes negatively buoyant, and accelerates downward to the surface. The IED, which is located along the sharp reflectivity gradient on the inner edge of the rainband, acts to sharpen this gradient by evaporating the rain as it falls.

The LLD begins as radial inflow approaching the rainband, located between 2- and 4-km altitudes. Once the airflow is beneath the heavily precipitating cell denoted by the 40-dBZ contour, precipitation drag forces the air downward and sustained evaporative cooling causes it to become negatively buoyant. Thus, the air accelerates downward. As a result of the downdrafts encountering the ocean surface, there is a region of divergence beneath the convective cell, which has consequences for the overall rainband. The airflow then continues radially inward along with the background vortex flow. In the diagram, both downdrafts overlap the overturning updraft to emphasize the three-dimensional nature of the airflow within the rainband.

As discussed by Hence and Houze (2008), the overturning updraft begins as radial inflow in the lower levels, where horizontal vorticity \(\xi\) is created by vertical wind shear. As this airflow rises, the horizontal vorticity \(\xi\) of the radial inflow is tilted, leading to positive vertical vorticity \((\zeta > 0)\) within the updraft core. In addition, the horizontal vorticity \(\xi\) of the updraft’s environment (i.e., the basic-state shear) is tilted to create a vertical vorticity couplet straddling the rotating updraft. One manifestation of the tilting of the environment vorticity is a negative vorticity \((\zeta < 0)\) anomaly located radially inward of the updraft. The positive vorticity characterizing the rising air is stretched and advected upward, constituting a positive vorticity flux into the middle levels. Vorticity flux per unit area is defined as \(F_z = w\zeta\). At around the 6-km level, \(\partial F_z/\partial z < 0\), which creates vertical convergence of positive vorticity and a local maximum of positive vorticity. The radially outer flank of this positive vorticity is manifested as the SHWM discussed by Hence and Houze (2008). The newly formed negative vorticity on the radially inner flank of the updraft becomes collocated with the IED, which then advects it downward. At around the 3-km level, \(\partial F_z/\partial z > 0\), which creates vertical convergence of negative vorticity and a local maximum of negative vorticity. The negative vorticity maximum increases the tangential winds just radially inward, creating or enhancing the LLWM. The LLWM, which is located within radial inflow as depicted in the conceptual model, is also a local maximum in angular momentum relative to the entire storm. An increase in inward flux of angular momentum at lower levels results, possibly strengthening the storm as a whole.

13. Conclusions

We have analyzed airborne Doppler radar observations collected in Hurricane Katrina (2005) by the NCAR ELDORA radar system to explore the physics and dynamics of the downdrafts associated with the
A convective/stratiform separation algorithm applied to the reflectivity data objectively identified the principal rainband’s main line of intense convection. By composite analysis of radial cross sections of the ELDORA data with respect to this objectively defined line of convection, we have identified the characteristics of the two types of downdrafts previously identified by Barnes et al. (1983) and Hence and Houze (2008). These two features, referred to as the low-level downdraft and the inner-edge downdraft, are not organized on the scale of the vortex, but rather occur on the convective scale, correlated spatially with the convective-scale updrafts.

The structure and behavior of the two downdrafts are summarized in Fig. 19. The LLD, which originates as radial inflow in the lower levels, is likely forced by precipitation drag and subsequent evaporative cooling. The IED, which originates from the hurricane’s radial outflow in the middle levels, is likely initially forced by pressure perturbation fields created in response to the buoyant overturning updrafts. In some cases, the air subsequently becomes negatively buoyant by evaporative cooling of the heavy precipitation within the rainband. Evaporation from the IED then apparently creates the sharp gradient of reflectivity along the inner edge of the principal rainband, which is a well-observed feature in this and past studies. We suggest that the vorticity dynamics of the overturning updraft and IED create a low-level wind maximum that lies in a region of inward radial flow, possibly strengthening the circulation of the overall storm when its angular momentum is advected inward by the low-level radial inflow of the tropical cyclone. We also suggest that the updraft and downdraft cores work together to sustain the convection in the principal rainband by forming new cells on its upwind end that are advected downwind, making the principal rainband quasi-stationary relative to the storm center. Through this interaction of convective-scale features, the principal rainband can continue to impact the overall storm, whether it strengthens the storm by generating and advecting potential vorticity to the center, or weakens the storm by inhibiting the inflow of warm, moist air (May and Holland 1999; Chen and Yau 2001; Franklin et al. 2006; Powell 1990b).
Future studies should employ a numerical model to test the hypotheses presented and described in this study. This model must be highly resolved in order to simulate convective-scale motions similar to those observed in the RAINEX dataset. Further investigation is needed of boundary layer processes below 500 m since the boundary layer was beyond the scope of the ELDORA dataset. Further investigation is also needed of outer rainbands for convective-scale comparisons with the rainbands in the inner core of the hurricane.

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APPENDIX A

Convective–Stratiform Separation Algorithm

The convective–stratiform separation algorithm using radar reflectivity data is based on ideas presented in Houze (1973) and was first defined by Churchill and Houze (1984). The separation technique first identifies convective regions based on the local peakedness of reflectivity values at a certain low-level altitude, and then designates the remaining pixels as stratiform. This is done because convective precipitation has a distinguishing radar signature as a tall tower of locally intense reflectivity values, while stratiform echoes are highly variable.

Individual pixels are convective centers if their reflectivity value exceeds the local background reflectivity (Z_{bg}) by at least a certain amount called the convective center criterion (ΔZ_{cc}; Steiner et al. 1995). Here Z_{bg} is defined as the average of nonzero and nonnegative radar reflectivity values within a radius of 11 km around the grid point. Pixels with reflectivity values greater than or equal to a certain threshold intensity (Z_{ti}) are also identified as convective centers. In alterting the algorithm for their study, Steiner et al. (1995) warned that differences in radar instrumentation affect the algorithm’s ability to identify radar signatures. Thus, the algorithm must be “tuned” for the particular radar collecting the data. Yuter and Houze (1997) introduced the convective center criterion (ΔZ_{cc}) as a cosine function of Z_{bg} given by

\[ ΔZ_{cc} = a \cos \left( \frac{1}{b} \frac{Z_{bg}}{2} \right), \]

where \( a \) and \( b \) are arbitrary parameters defined by the user, making tuning simple and intuitive. Once convective centers are identified, the pixels within a certain radius of the convective center, called the convective radius (R), are also classified as convective. Here R is a piecewise function depending on \( Z_{bg} \) given by

\[ R = \begin{cases} 0.5, & Z_{bg} < 20 \\ 0.5 + 3.5 \left( \frac{Z_{bg} - 20}{15} \right), & 20 \leq Z_{bg} < 35 \\ 4, & Z_{bg} \geq 35 \end{cases} \]

where \( R \) is in units of kilometers. This function remains similar to the \( R \) from previous studies in that the variance in radii is applied over a \( Z_{bg} \) range of 15 dBZ; however, the range of possible \( R \) is altered to better match the higher resolution of the current dataset. The remaining pixels with reflectivity less than a certain threshold \( Z_{we} \) are classified as “weak echo.” Weak echoes are ambiguous as to whether they are convective or stratiform. All of the remaining unclassified pixels are defined as “stratiform.”

In the tuning of the algorithm, the variables \( a, b, Z_{ti}, \) and \( Z_{we} \) were chosen such that the objective classification of the low-level reflectivity data was most consistent with a subjective classification of precipitation type based on vertical cross sections of the reflectivity data. In addition to meeting the previously mentioned criteria, the \( Z_{bg} \) calculated at an individual pixel must be greater than or equal to another background reflectivity value, \( Z_{bg2} \), defined as the average reflectivity within the same radius as \( Z_{bg} \) but at the approximate altitude of the bright band, which in this case was 4.5 km. The best classification results were found when the algorithm was run at the low-level altitude of 2 km, and the parameters were \( a = 9, b = 45, Z_{ti} = 42 \) dBZ, and \( Z_{we} = 20 \) dBZ.

APPENDIX B

Thermodynamic Retrieval

Using the equations of motion, thermodynamic equation, and water continuity equations, horizontal and vertical gradients of temperature and pressure perturbations can be deduced from the available reflectivity and wind fields (Gal-Chen 1978; Houze 1993, chapter 4). The formulation used for this analysis follows the approach described by Roux and Ju (1990) and Roux et al. (1993). In this formulation, anelastic mass continuity is used as a constraint and solutions are obtained by a variational...
technique that minimizes errors over the entire spatial domain. Since the current dataset is assumed to be instantaneous, the retrieval was done in the steady state. The thermodynamic retrieval was performed on the three-dimensional data volume between 25.48°–26.38°N and 87.49°–87.89°W. Perturbations were calculated relative to the upper-air sounding at Tallahassee, Florida, at 1800 UTC 28 August 2005. To prevent computational errors in the retrieval equations, we applied a three-step filter (Leise 1982) prior to the retrieval, limiting the resolved scales to 3.2-km wavelength. Horizontal and vertical velocity values were limited by a minimum threshold magnitude of 1 and 0.5 m s⁻¹, respectively.

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