A Numerical Study of a Rotating Downburst

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ABSTRACT

Previous studies have revealed that convective storms often contain intense small-scale downdrafts, termed "downbursts," that are a significant hazard to aviation. These downbursts sometimes possess strong rotation about their vertical axis in the lower and middle levels of the storm, but studies of how this rotation is produced and how it impacts downdraft strength are lacking. In this study a three-dimensional cloud model was used to simulate a rotating downburst based on conditions observed on a day that produced rotating downbursts. It was found that rotating downbursts may occur when the direction of the wind shear vector in the middle levels of the troposphere varies with height. In the early stages of the convective system, vertical vorticity is generated from tilting of the ambient vertical shear by the updraft, resulting in a vertical vorticity couplet on the flanks of the updraft. Later, the negative buoyancy associated with precipitation loading causes the updraft to collapse and to be eventually replaced by a downdraft downshear of the midlevel updraft. When the direction of the vertical shear vector varies with height, a correlation may develop between the location of the vertical vorticity previously produced by the updraft at midlevels and the location of the developing downdraft. This mechanism causes downbursts to rotate cyclonically when the vertical shear vector veers with height and to rotate anticyclonically when the vertical shear vector backs with height. The rotation associated with the downburst, however, does not significantly enhance the peak downdraft magnitude. The mechanism for the generation of vorticity in a downburst is different from that found for supercell downdrafts, and, for a given vertical shear vector, downbursts and supercell downdrafts will rotate in the opposite sense.

1. Introduction

It has long been known that downdrafts occur with convective cells. For example, Byers and Braham (1949) documented the existence of downdrafts in convective storms that approached magnitudes of \(-20\) m s\(^{-1}\). Fujita and Byers (1977) more recently concluded that convective downdrafts can still be intense near the surface, after observing a downdraft of nearly \(-4\) m s\(^{-1}\) at a height of only 100 m above ground level. These intense low-level downdrafts, termed downbursts, have been shown to pose a hazard to aircraft upon takeoff or landing (Fujita and Caracena 1977). A number of studies (e.g., Forbes and Wakimoto 1983; Fujita and Wakimoto 1983; Fujita 1985; Roberts and Wilson 1984; Hjelmfelt 1987; Kessinger et al. 1988; Mahoney and Elmore 1991) have found that the periphery of downbursts are sometimes characterized by intense horizontal shear and/or distinct rotation centers. In one example, Kessinger et al. (1988) used dual-Doppler analysis to investigate numerous downbursts that were collocated with small-scale (kilometer-scale) cyclones.\(^1\) The downdrafts were typically associated with cyclonic vorticity, although one mesoanticyclonic center was observed. The magnitude of the vorticity in one of the mesocyclones in this storm approached 0.1 s\(^{-1}\) (Fig. 1), which is comparable to the vorticity present in the parent mesocyclone associated with a tornado (Kessinger et al. 1988).

A number of authors (Emanuel 1981; Wolfson 1983; Fujita 1985) have hypothesized that the rotation associated with downburst mesocyclones may actually increase downdraft magnitude. In contrast, Kessinger et al. (1988) used thermodynamic data retrieved from the equation of motions and Doppler radar–derived winds to show that the vertical pressure force associated with the rotation actually tended to oppose the downward acceleration below cloud base. Kessinger et al. (1988) did speculate, however, that the vertical pressure forces associated with the rotation may enhance the downdraft magnitude above cloud base. Numerical studies by Proctor (1989) and Trapp and Das (1990) have also examined the effect of rotation on downdraft intensity. These studies produced the rotation by either 1) concentrating the ambient vertical vorticity through

\(^{1}\) In our study these small-scale cyclones will be called mesocyclones, but our use of this term is not meant to imply that these cyclones are identical to the mesocyclones associated with supercell convection.
we present simulations with the model initialized using various hodographs in an effort to generalize our results. We determined that the changes in the direction of the vertical shear of the horizontal wind (hereafter called vertical shear) in the middle levels of the ambient environment dictate the type of rotation associated with a downburst. At first, this result may seem to be similar to the creation of rotation in a supercell convective system (see Rotunno and Klemp 1985 for a review). However, the mechanisms that concentrate the vorticity in the two cases are different, and for a given shear vector, downbursts and supercell downdrafts will rotate in the opposite sense.

2. Model characteristics and initial conditions

The numerical model used in this study is a version of the three-dimensional cloud model described by Klemp and Wilhelmson (1978). The model uses the complete set of compressible equations with a subgrid kinetic energy equation. We choose not to include the Coriolis terms. Our version of the model employed a modified Kessler (1969) parameterization for the microphysical processes. The parameterization used in this study did not include ice processes. The accompanying observational study of Kessinger et al. (1988) and the idealized numerical simulations of this system by Proctor (1988, 1989) suggest that ice processes are secondary to evaporation in maintaining downbursts on this day. Further details on the impact of neglecting ice processes in simulations of storms with low-level downdrafts can be found in Knupp (1985).

The simulation used in these experiments covered 30 km × 30 km × 16 km with a mesh size of 500 m in each horizontal direction and a smoothly varying stretched interval in the vertical ranging from 246 m between the lowest levels to a 734-m interval at the top of the domain. The simulated convective storms are initialized by placing an axially symmetric thermal perturbation in an environment that is initially horizontally uniform. The thermal was centered at 1400 m above the surface and decreased in the horizontal from a maximum of 2.5°C at the center to 0°C at the edge. This type of initialization allows us to study the full life cycle of a convective storm in contrast to more idealized numerical experiments that have generated the downdraft phase of the system only through placing a precipitation core within the ambient environment (i.e., Proctor 1988, 1989). Our use of a full three-dimensional simulation also allows us to address the effects of rotation and a turning vertical shear vector on downburst dynamics. In contrast, previous idealized simulations of this system using axisymmetric models (i.e., Proctor 1988, 1989) were unable to obtain these effects and thus missed certain features of the observed storm, such as the rotation centers and an asymmetric storm outflow.
FIG. 2. A skew T-logp plot of temperature and dewpoint temperature used to initialize the simulation. The profiles are based on a sounding launched at 1800 LDT from Denver on 30 July 1982 with slight modification. These modifications include smoothing due to interpolation to the model vertical grid points, further smoothing of the temperature lapse rate in the lower boundary layer, and increasing the water vapor content at second model grid point by approximately 30% so that convection was able to be initiated. This 1800 LDT sounding is considered to be representative of the environment just ahead of the observed rotating downbursts. The shaded area is approximately proportional to the amount of energy a parcel lifted with the mean characteristics of the lowest 50 mb would have if it ascended undiluted through the undisturbed environment.

The vertical profiles of temperature and moisture used in the experiment are shown in Fig. 2 with the vertical profile of equivalent potential temperature $\theta_e$ presented in Fig. 3. These profiles are based on a sounding taken in the experimental area at Denver at 1800 LDT 30 July 1982. The downburst studied in Kessinger et al. (1988) occurred within an hour of this sounding launch. The profile of $\theta_e$ depicts a convectively unstable atmosphere with a broad minimum in $\theta_e$ located between 3 and 6 km. The magnitude of the convective available potential energy (CAPE) for a representative lower-level parcel rising vertically through this undisturbed sounding is approximately 2300 m$^2$ s$^{-2}$. For comparison, Weisman and Klemp (1984) suggest that the CAPE for moderately unstable cases ranges between 1500 and 2500 m$^2$ s$^{-2}$. The temperature lapse rate observed by this sounding was nearly dry adiabatic up to approximately 640 mb. The deep dry-adiabatic lapse rate has been previously shown to be associated with strong low-level downdrafts (Krumm 1954; Caracena et al. 1983; Wakimoto 1985).

The wind hodograph used in this experiment is shown in Fig. 4. The sounding is generally characterized by weak shear below 4.5 km in height. The bulk Richardson number for this sounding is approximately 244. Steady supercell storms are predicted to occur with bulk Richardson numbers less than 40 (Weisman and

FIG. 3. The initial vertical profile of $\theta_e$ used in the simulation. The thermodynamic profile was based upon the sounding (Fig. 2) taken from Denver at 1800 LDT 30 June 1982.

FIG. 4. Hodograph of the winds determined from the 1800 LDT Denver sounding (Fig. 2). The heights (km) of the wind measurements are indicated along the hodograph curve. The dashed arrow represents the approximate speed of the simulated convective cell averaged over its lifetime.
Klemp 1984) so that the simulated storm should be decidedly nonsteady in nature. The combination of the high bulk Richardson number together with the magnitude of the buoyant energy and the deep low-level dry-adiabatic layer suggests intense but nonsteady convective systems with strong downdrafts in the lower levels. Since the convective cells observed on this day had these characteristics, some added confidence can be gained in the belief that the sounding is adequately representative of the environmental conditions. However, the reader should keep in mind that although we reproduce many aspects of the observed storms, our primary interest in this study is to isolate the mechanism that produces the rotation in these storms. For this reason we have chosen a relatively simple set of experiments (i.e., no ice phase, no surface fluxes, a uniform environment, and warm bubble initialization).
3. Overview of the simulated storm

The life cycle of the simulated convective cell is similar to that described for the evolution of air mass (or unsteady) thunderstorms in observational studies as early as the Thundestorm Project (Byers and Braham 1949) and replicated in early numerical simulations (e.g., Orville and Sloan 1970). In particular, the evolution of the convective cell takes place in 45 min as the cloud varies from an intense updraft, to a mixed updraft–downdraft system, and finally to a weak broad downdraft. A series of vertical cross sections of precipitation, cloud water, and velocity vectors through the simulated storm (Fig. 5) reveals crucial aspects of the system’s life cycle. The maximum updraft of 38 m s⁻¹ took place at 1500 s into the simulation. At this time, the convective cloud was dominated by updrafts with a general absence of a downdraft below cloud base and weak downdrafts present on the middle and upper flanks of the cloud (Fig. 5a). The precipitation field (Fig. 5a) shows that significant precipitation loading occurs in the middle layers of the storm. Between 1500 and 1800 s, cloud top increased from 8.5 to 12 km in height and the cloud developed a distinct anvil structure (Fig. 5b). The precipitation loading aloft also increased significantly with the bulk of the convective updraft still coinciding with a maximum in the precipitation loading.

At 2100 s (Fig. 5c), a low-level downdraft formed as heavy precipitation fell from cloud base. The vertical cross section of velocity vectors and precipitation at this time suggests that the updraft became cut off from its source of unstable low-level air as a nearly 1-km rise in cloud base took place between 1800 and 2100 s. The vertical cross section at 2400 s (Fig. 5d) shows that cloud base continued to rise as light precipitation reached the surface and heavy precipitation continued to fall from cloud base. The reduction in the area covered by heavy precipitation from 1800 to 2400 s is consistent with the strong evaporation that we will later show has taken place during this time. From examination of the flow in Figs. 5c and 5d and trajectory analysis, one will later see that the circulation into the downdraft originates from between 4 and 5 km in height. This zone lies within the broad minimum in $\theta_e$ (Fig. 3).

The peak downdraft took place near 2550 s with magnitudes of nearly 13 m s⁻¹ (Fig. 6). The downdraft at this time was associated with 1) a decrease in precipitation content with height below 2.5 km, 2) the maximum low-level divergence found in the storm ($\sim 2.2 \times 10^{-2}$ s⁻¹), 3) an asymmetric surface outflow, and 4) centers of rotation about a horizontal axis (typically termed rotors) near the leading edge of the gust front. After the period of strongest downdrafts, the cloud became characterized by a weakening and broadening of the downdraft and a continuation of the asymmetric outflow of cold air at the surface (Fig. 5e). The strongest downdraft in the later stages of the downburst occurred near the leading edge of the spreading cold pool (Fig. 5f) in association with very light precipitation and one of the previously mentioned rotors.

Horizontal cross sections of relative winds, vertical motions, and precipitation are shown for the surface, 3 km, and 6 km in height at 2100 and 2700 s (Fig. 7), revealing the horizontal circulation features within the storm. At 2100 s, the flow at 6 km suggests a vertical vorticity couplet in the vicinity of the strong updraft (Fig. 7a). The updraft is still relatively strong at 6 km,
but there is evidence of precipitation falling and a downdraft developing downshear of the updraft. At this time, the flow at the surface is relatively undisturbed and the flow at 3 km is convergent. At 2700 s, there is strong divergence evident at the surface in association with the impact of the downburst (Fig. 7b). While at this time the vorticity couplet and updraft are no longer clearly evident at 6 km, there is a mesocyclone on the flank of the strong downdraft at 3 km and some cyclonic rotation at 6 km. The time history of the vertical vorticity reveals that the peak rotation associated with the mesocyclone occurred at a height of 2.5 km at approximately 2700 s into the simulation. Since the peak downdraft took place just above 1 km in height at 2550 s into the simulation, it is evident that the strongest rotation lags the peak downdraft in both time and height.

In many regards, the downdraft magnitude, cell life cycle, rotor circulation, and mesocyclone center are similar to the Doppler radar observations presented in the Kessinger et al. (1988) study. For example, the peak Doppler-derived downdrafts on this day ranged between −11 and −15 m s⁻¹, while the downdraft in the simulated system was nearly −13 m s⁻¹. The maximum divergence in the simulation of approximately 2.2 × 10⁻² was also close to the observed values. In addition, both the simulated and observed storm contained a mesocyclone (Figs. 1 and 7). Despite these similarities, the comparison should be treated with some caution. For example, the observed convective storm was composed of many cells, while in our simulation we are initializing the model and then studying the life cycle of a single convective cell. Our initialization used the smallest thermal perturbation that resulted in a realistic appearing convective cell. The general qualitative aspects of the systems, such as the rotation centers and strong downdraft below cloud base, were not changed by initializing the model with something stronger thermals. The actual magnitudes of the simulated updraft and downdraft, however, were dependent upon the magnitude of the initial thermal perturbation making quantitative agreement between the observed and simulated downdraft somewhat fortuitous. The lack of an ice phase in the simulation also prevents a quantitative comparison between the radar observations and the simulated system.

While we have previously mentioned that the convective life cycle is similar to classical conceptual models of unsteady convection, the presence of the mesocyclone in the simulations and observed storm together with the rapid acceleration of the downdraft below cloud base and the pronounced rotor represents significant differences from the classical model. These

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**Fig. 6.** An east–west vertical cross section through the maximum downdraft at 2550 s into the simulation. A vector length of one grid interval denotes a wind of approximately 20 m s⁻¹. The light and dark shading defines the region covered by rainfall and the 2.25 g kg⁻¹ rainwater mixing, respectively. The coordinate system is located at y = 12 km and x = 8 to 18 km within the larger model domain.

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**Fig. 7.** Horizontal cross sections of relative wind at 0, 3, and 6 km in height for (a) 2100 s and (b) 2700 s into the simulation. A vector length of one grid interval represents a wind of 8 m s⁻¹ at the 3- and 6-km height, while because of the strong winds associated with the outflow, a single grid interval represents a 20 m s⁻¹ at the surface. The vertical velocity field is superimposed and contoured in intervals of 2 m s⁻¹ and rainwater contents in excess of 0.0005 and 0.006 g kg⁻¹ are lightly and darkly stippled.
features are particularly important for aviation. For example, the danger from strong low-level downdrafts is well established and the rotor circulations have been implicated as a possible source of wind shear in the Dallas–Fort Worth air crash (Fujita 1985). The possible danger to aviation from the mesocyclone is less clear. While strong, the crosswinds associated with the rotation are likely to increase pilot response time to a downburst and perhaps also increase the chance for pilot error; we will later show that the rotation is generally small in the lower levels, where the danger to aviation operations is greatest.

4. Analysis of the simulated storm

In this section we will analyze the simulated storm system to determine how the downburst is generated and show how the vorticity associated with the mesocyclone is created and destroyed. We will also address the impact of the rotation on downburst magnitude.

a. Downburst forcing

The anelastic momentum equations can be written as

\[
\frac{dv}{dt} = \frac{\partial v}{\partial t} + v \cdot \nabla v = -c_p \rho_0 \nabla \pi + Bk + F, \tag{1}
\]

where \(v\) is the vector velocity \((u, v, w)\) in Cartesian coordinates, \(c_p\) the specific heat of dry air at constant pressure, \(\theta_0\) the mean virtual potential temperature, \(\pi\) the perturbation Exner function for pressure, \(B\) the total buoyancy, and \(F\) the turbulent mixing. The formation of the strong downdraft can be addressed through examining the vertical component of (1) written as

\[
\frac{dw}{dt} = -c_p \rho_0 \frac{\partial \pi}{\partial z} + B + F_z. \tag{2}
\]

The primary terms driving the vertical accelerations are the buoyancy and the force resulting from vertical gradients in pressure. The buoyancy term \(B\) can be approximated as

\[
B = g \left( \frac{\theta'}{\theta} + 0.61 q_v' - q_b - q_r \right) \tag{3}
\]

with \(g\) the acceleration due to gravity, \(\theta\) the potential temperature, and \(q_v, q_b, q_r\) the mixing ratios of water vapor, cloud water, and rainwater, respectively. The overbar and prime designate mean and deviation states, respectively.

The role of rotation on the vertical pressure gradient in (2) can be assessed by writing a diagnostic equation for pressure. This procedure has been found useful for analyzing flows in simulated convective storms (e.g., Williamson and Ogura 1972; Schlesinger 1980; Rotunno and Klemp 1982, 1985; Klemp and Rotunno 1983) and in radar observations of downbursts (Parsons and Kropfli 1990). A diagnostic equation for pressure can be obtained by taking the divergence of (1), substituting the anelastic continuity equation, \(\nabla \cdot (\bar{\rho} v) = 0\), and ignoring the mixing terms to obtain

\[
\nabla \cdot (c_p \bar{\rho} \bar{v} \nabla \pi) = -\nabla \cdot (\bar{\rho} v \cdot \nabla \pi) + \frac{\partial \bar{\rho} B}{\partial z}, \tag{4}
\]

where \(\rho\) is the density. Since (4) is linear in pressure, the various contributions to the pressure field can be separately assessed and (4) can be rewritten as

\[
\nabla \cdot (c_p \rho \bar{v} \nabla \pi) = -2\rho \left( \frac{\partial v}{\partial x} \frac{\partial u}{\partial y} + \frac{\partial v}{\partial y} \frac{\partial w}{\partial y} \right) \tag{A}
\]

\[
-2\rho \left( \frac{\partial u}{\partial z} \frac{\partial w}{\partial x} + \frac{\partial w}{\partial z} \frac{\partial w}{\partial y} \right) \tag{B}
\]

\[
-\rho \left[ \left( \frac{\partial u}{\partial x} \right)^2 + \left( \frac{\partial v}{\partial y} \right)^2 + \left( \frac{\partial w}{\partial z} \right)^2 - \frac{d^2 \ln \bar{\rho}}{dz^2} \right] \tag{C}
\]

\[
+ \frac{\partial \bar{\rho} B}{\partial z} \tag{D}
\]

where \(A\) and \(B\) are the contribution to the pressure field from fluid shear, \(C\) is the contribution from fluid extension, and \(D\) is the contribution to the pressure field from vertical gradients in buoyancy. Terms \(A\), \(B\), and \(C\) are commonly called the dynamic contributions to the pressure since they are associated only with the wind field. Term \(A\) can be visualized as representing the effect of rotation about a vertical axis, while \(B\) is the contribution due to the interaction between the updraft and the vertical shear (effects of rotation about the horizontal axes). The method of solving (5) and obtaining estimates of the contributions to the vertical pressure force is described in Rotunno and Klemp (1985). If the strong downdraft is partly driven by pressure forces associated with rotation, then there should be a significant contribution to the vertical acceleration from the dynamic pressure force in (4) and specifically a favorable vertical distribution of pressure associated with term \(A\) in (5).

In order to determine whether the downburst was driven by the generation of negative buoyancy or by vertical pressure forces induced by rotation, the contributions to the vertical acceleration in (2) due to the buoyancy and dynamic portion of the vertical pressure force were calculated along a time-dependent trajectory passing through the peak downdraft in Fig. 6. The parcel in this trajectory first rose from a height of approximately 4.2 to 4.5 km, slowly descended to 4 km, and then descended rapidly after 2100 s (Fig. 8). Examination of the buoyancy and the dynamic contribution
results are consistent with the retrieval results presented in Kessinger et al. (1988) that show a significant portion of the precipitation must evaporate to account for the temperature deficit observed below cloud base, but that precipitation loading initially triggers the downdraft.

While the comparison is favorable, one must recall that our simulation does not have an ice phase. While this suggests that comparisons between our findings and atmospheric flows must be treated with some caution, one can also conclude that in this environment, a rain-only simulation produces a reasonable rendition of a downdraft with a downdraft magnitude that is within 20% of the observed. The effects of an ice phase on downdraft intensity have been previously investigated and are discussed, for example, in Knupp (1985), Srivastava (1987), and Proctor (1988, 1989). Both Proctor (1989) and Srivastava (1987) argue that an ice phase is more crucial to producing a downdraft when more stable boundary layers are present by delaying the cooling to lower levels where it is more effective in producing strong surface divergence. In this case, the lapse rate is rather steep particularly below 650 mb.

Our findings show that the downdraft is primarily driven by negative buoyancy and not driven by a dynamic pressure force. This conclusion is similar to studies of strong downdrafts in nonrotating storms (e.g., Knupp 1989). Further evidence for the hypothesis that the strong downdrafts are driven by negative buoyancy and not a favorable vertical pressure gradient associated with the mesocyclone can be found through

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**Fig. 8.** Vertical motion (solid line), the dynamic contribution to the vertical pressure gradient [terms A–C in (5) and denoted by a dotted line], and buoyancy (dashed line) for a time-dependent trajectory through the maximum downdraft in Fig. 6. The height of the parcel is indicated along the vertical motion curve.

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**Fig. 9.** Contributions to the buoyancy from the precipitation loading (solid line), potential temperature deviation (dashed line), and perturbations in the water vapor mixing ratio (dotted line) terms in (3) for the same trajectory as in Fig. 8.
examining the vertical pressure force associated with term A in (5). This term, shown in Fig. 10, indicates that rotation results in a weak upward-directed pressure force that opposes the peak downdraft. This result contrasts with previous studies (e.g., Wolfson 1983; Roberts and Wilson 1989) that hypothesized that rotation may actually cause downbursts due to vertical pressure forces. Our findings, however, are consistent with the thermodynamic retrieval analysis undertaken from radar observations of this case (Kessinger et al. 1988). The explanation for the upward-directed pressure force in the observations (Kessinger et al. 1988) was that 1) the rotation center was associated with reduced pressure [see (5)] and 2) the rotation center and its associated lower pressure were located above the peak downdraft so that the vertical pressure gradient force was directed upward near the height of the peak downdraft. A similar situation occurs in the simulations as the pressure reduction associated with the rotation (Fig. 11) was also located above the peak downdraft. In addition, as shown in Fig. 8, a far larger upward-directed pressure force occurs as the downdrafts near the surface. This upward-directed pressure force is dynamic in nature and due to strong divergence [term C in (5)] resulting from the downdraft impacting the surface (Fig. 12).

b. Formation of the mesocyclone

In order to further understand the relationship between the rotation and the downdraft, we examined the processes producing vorticity in the mesocyclone using time-dependent trajectories constructed from model output. To the accuracy of the Boussinesq approximation, an equation for the time rate of change of the vertical vorticity, \( \zeta \), can be obtained by taking the curl of (1) and organizing the terms so that

\[
\frac{d\zeta}{dt} = \omega_H \cdot \nabla H w + \zeta \frac{\partial w}{\partial z} + F'_{HH},
\]

(6)

with \( \omega_H \) the horizontal vorticity (\( \nabla \times V \)) and \( F'_{HH} \) the mixing term. The first term on the right-hand side of (6) is the so-called tilting term and the second term is the stretching term. In our analysis we found the mixing term to be insignificant, except in the lowest levels where it contributes slightly to the dissipation of the mesocyclone. For this reason we have not plotted the dissipation term in our figures.

The vertical motion, vertical component of vorticity, and height are indicated in Fig. 13a for a trajectory calculated backward in time from the center of the mesocyclone near cloud base. From Fig. 13a, it is evident that the parcel subsides slightly and then begins
VORTICITY PRODUCTION IN THE MESOCYCLONE

Fig. 13. (a) Plots of vorticity (solid line) and vertical motion (dashed line) versus time for points along a time-dependent trajectory backward in time from a location at the center of the mesocyclone. The height of the parcel along the trajectory is indicated along the vorticity curve. In a coordinate system moving with the convective cell, the horizontal origin of the parcel was east-northeast of the mesocyclone. (b) As in (a) but for vorticity generation by the tilting (solid) and stretching terms (dashed) in (6).

being retarded by the vertical pressure force associated with rotation. For example, the parcel associated with the time-dependent trajectory through the maximum rotation (Fig. 13) always lies above the parcel that passes through the strongest downdraft (Fig. 10) so that the low pressure induced by the rotation lies above the peak downdraft, which as we stated earlier would have a retarding effect on the strong impulsive downdraft that is the feature of practical interest for the study of downbursts.

The corresponding trajectory forward in time from the mesocyclone center at two different times allows us to address the demise of the mesocyclone. The stretching and tilting terms for the two trajectories (Fig. 14) show that as the parcel descends, the stretching term decreases rapidly from its large positive value and becomes small. Realizing that the peak downdraft is located below rotation center, this finding can be explained by the peak downdraft decelerating as it approaches the surface, which reduces the vertical gradient of \( \omega \) in the stretching term in (2). While the stretching term becomes negligible, the vorticity is rapidly decreased by a negative tilting term (Fig. 14). The negative tilting term is associated with an increase in vertical shear due to the development of the outflow in the lower layers and the previously presented observation that the rotation center lies on the flank of the downdraft (e.g., Fig. 7).

VORTICITY DISSIPATION IN MESOCYCLONE

Fig. 14. As in Fig. 13b but for two trajectories forward from the mesocyclone center at 2550 and 2700 s into the simulation.
5. Generalizing the vorticity production mechanism

In our simulation based on the 30 June 1982 sounding, we were able to reproduce a rotating downburst. The vertical vorticity associated with the mesocyclone originally formed due to tilting of the environmental vertical shear by the updraft and subsequently increased via stretching (and to a lesser degree tilting) as the updraft collapsed and a precipitation-induced downdraft developed in this unsteady storm. After our analysis of this storm, however, the questions of how the vorticity and the downdraft become correlated and why cyclonic vorticity is favored are left unanswered. As a first step toward answering these questions, we undertook simulations with the identical thermodynamic sounding, but with simplified wind hodographs. These experiments show that a turning vertical shear vector in the middle levels of the environment (\( \sim 3-10 \) km) was necessary to produce the mesocyclone. The mesocyclone from one of these simulations, which had no shear either below 3.5 km or above 12 km in height but included vertical shear in the layer between 3.5 and 12 km, is shown in Fig. 15. The vertical shear vector in this simplified hodograph veers with height through its entire depth (Fig. 16). We will follow the evolution of this storm in the context of some simple dynamic concepts to define the mechanism for the creation of the mesocyclone.

For a homogeneous fluid, Rotunno and Klemp (1982) were able to linearize the shallow, inviscid anelastic equations of Ogura and Phillips (1962) to show that

\[
\frac{D}{Dt} \zeta = k \cdot \left( \frac{dV}{dz} \times \nabla w' \right),
\]

where \( \zeta \) is the vertical component of the vorticity now defined as \( \zeta = \partial v / \partial x - \partial u / \partial y \). Equation (7) can be interpreted as the creation of vertical vorticity from the effect of tilting of the vortex lines with the vertical vorticity perturbations oriented perpendicular to the vertical shear vector. From (7) one can see that for an axisymmetric updraft, cyclonic (anticyclonic) vorticity will be created on the right (left) side of the shear vector (Fig. 17c). While this relationship was derived from linearization, this spatial distribution also holds as the vorticity processes becomes nonlinear (Davies-Jones 1984; Rotunno and Klemp 1985).

The vorticity pattern relative to the updraft for early in the simulation with the simplified wind hodograph (Fig. 18) shows a vorticity pattern similar to that predicted by linear theory (Fig. 17c) with cyclonic (anticyclonic) vorticity to the right (left) of the updraft. The position of the vorticity maxima and minima together with the relative flow, also shown in Fig. 18, suggests that there is some advection of vorticity downwind from the initial peaks. The vorticity production due to the tilting in (7) is symmetric and does not result in a bias toward either sign as long as the vertical motion pattern remains symmetric with time.

Having shown that the initial vorticity generation...
fits a well-established conceptual model of the updraft interacting with the ambient vertical shear, we next will show how the correlation occurs between the precipitation induced downdraft and the vorticity. The location of the falling precipitation relative to the updraft-generated vorticity pattern (Figs. 19a and 20a) shows that the precipitation amounts are correlated with cyclonic vorticity. At later times the location of the downdraft, which is initially driven by precipitation loading, is also correlated with cyclonic vorticity (Figs. 19b and 20b). These results can be easily explained by examination of the hodograph (Fig. 16) and the three-dimensional perspective (Figs. 19a and 19b), which show that the precipitation is carried downshear by the stronger winds aloft so that as it falls, it passes through the cyclonic side of the vorticity pattern created by the updraft in the middle levels of the storm. (From Fig. 7 we can also see that this process also took place in the control simulation.) As intensification of the downdraft takes place in the lower levels (Fig. 19b), the vorticity is stretched and eventually a mesocyclone forms. Since there is a spatial displacement between the locations of the developing downdraft core and the vorticity center, the stretching term is maximized on the flank of the downdraft (Fig. 21).

A schematic of the process generating the vorticity

**FIG. 17.** Schematic illustrating the behavior of the vorticity and pressure field described by (7) and (8) for vertical motions interacting with a vertically sheared mean flow. The hodographs are illustrated at left. The locations of the maxima and minima in pressure and vertical vorticity are described by H and L and by "+" and "−", respectively. The vertical pressure forces are denoted by a dark arrow, and the updrafts by a light arrow with a dark tip. (a) A downdraft with a straight hodograph with no correlation between vertical vorticity and the vertical pressure gradient. (b) A downdraft with a veering hodograph where a correlation develops between anticyclonic vorticity and a downward-directed pressure force. (c) An updraft with a veering hodograph where a correlation develops between cyclonic vorticity and an upward-directed pressure force and a downward-directed force and anticyclonic vorticity. Fig. 17c has been adapted from Rotunno and Klemp (1982).

**FIG. 18.** A horizontal cross section of relative flow vectors at 4.5 km in height for 1650 s into the simulation. A vector length of one grid interval denotes a wind of approximately 7.5 m s⁻¹. Updrafts stronger than 15 m s⁻¹ are stippled. The vorticity is contoured in intervals of 1.5 × 10⁻³ s⁻¹ with shading indicating cyclonic vorticity. The domain is located at x = 25−35 km and y = 30−45 km in the larger model domain.
in the mesocyclone is shown in Fig. 22. From the previous discussion, the generation of vorticity in the downdraft mesocyclone can be described simply as 1) the generation of vorticity by the updraft tilting the ambient vertical shear, 2) the formation of a downdraft downshear from the vertical shear vector in the middle and upper levels of the storm, and 3) the stretching of the vorticity as a strong downdraft develops in the lower levels. The tilting of the vertical shear by the downdraft is secondary but plays a role in determining the position of the vorticity maxima. In this mechanism, how the shear vector turns with height determines where the precipitation falls and whether it becomes correlated with the one side of previously created vorticity couplet.

In order to illustrate this hypothesis, simulations were undertaken with the following hodographs (Fig. 23): 1) a shear vector that increased with height but was unidirectional, 2) a half-circle shear vector that veered with height, and 3) a half-circle shear that backed with height. These three simulations resulted in 1) a downdraft with a vortex couplet, 2) a downdraft mesocyclone, and 3) a downdraft mesoanticyclone, respectively. This result is consistent with our hypothesis. For example, for a straight hodograph, there should be no preferred correlation between the precipitation-induced downdraft and the vorticity, while reversing the sign of the turning shear vector should reverse the sign of the rotation as the precipitation falls on the opposite flank of the initial updraft. In the two cases with strong rotation (Figs. 23b and 23c), the peak subcloud downdrafts of approximately 10.5 m s\(^{-1}\) are noticeably weaker than the 14.9 m s\(^{-1}\) downdraft found in the linear shear simulation (Fig. 23a).

6. Comparison to supercell systems

The notion that the sense of the rotation depends on the vertical shear vector found in this study is reminiscent of work on supercell convection. For example, Rotunno and Klemp (1982) showed that when the vertical shear vector turns with height, rotating up- and downdrafts result from a correlation between vertical pressure gradients associated with the interaction of \(w\) with the ambient vertical shear and the vorticity pattern created by tilting of the ambient vertical shear by \(w\).

We will first show that the supercell mechanism is different from the process found in this study. According to Rotunno and Klemp (1982), the pressure perturbations \(\pi'\) in the middle levels of the storm can be approximated by

\[
\pi' \sim \frac{dV}{dz} \cdot \nabla w'.
\]

(8)

For an axisymmetric downdraft, the relationship in (8) dictates that pressure deviations are aligned along the shear vector in contrast to the vorticity production terms. The high pressure areas are downshear from the downdraft, while the low pressure is upshear (Figs. 17a,b). For the case of a straight hodograph, the pressure forcing and vorticity production remain 90° out of phase and there is no bias in the vorticity production (Fig. 17a). When the wind shear changes with height, however, vertical gradients of pressure and thus vertical accelerations can occur in regions of significant vorticity production. For example, if the wind shear vector veers with height, the updraft and cyclonic vorticity production become correlated (Fig. 17c) and con-
versely, the downdraft is increased in the region where anticyclonic vorticity is produced (Fig. 17b). Hence, in this mechanism, the presence of a veering wind shear vector produces a tendency for cyclonically rotating

Fig. 21. A horizontal cross section of relative flow at 4.5 km in height for 2550 s into the simulation. A vector length of one grid interval denotes a wind of approximately 7.5 m s⁻¹. Downdrafts stronger than –5 m s⁻¹ are shaded and the stretching term in the vorticity equation is contoured in intervals of 1.0 × 10⁻⁴ s⁻². The domain is located at x = 20–35 km and y = 32–47 km within the larger model domain.

Fig. 22. Schematic illustrating how the vorticity becomes associated with the downdraft for a wind shear vector that veers with height. The wind profile is equivalent to the half-circle shear (as in Fig. 23b) with low-level easterlies, midlevel southerlies, and upper-level westerlies. A vorticity couplet (cyclonic vorticity specified by the “+”) is created by the updraft tilting the ambient vertical shear (denoted by A). The precipitation falls from downshear at higher levels into the cyclonic vorticity side of updraft (denoted by B). As a downdraft develops and later intensifies in the lower levels due to evaporation the cyclonic vorticity is enhanced by stretching (denoted by C).
updrafts and anticyclonically rotating downdrafts. In contrast, our simulation with a veering profile (Fig. 16) resulted in a cyclonic downdraft (Fig. 15), which is in the opposite sense to that predicted by the supercell mechanism. The rejection of the supercell mechanism is consistent with both the finding that our simulated cells are short-lived and therefore quite different from supercells, and that the downburst storm has a higher bulk Richardson number which suggests that buoyancy forces should dominate over dynamically induced vertical gradients of pressure. The domination of buoyancy forces in the downburst cell causes the updraft to collapse, which results in an interaction between the downdraft and the vorticity previously created by the updraft. This interaction does not occur in supercells where the vertical motion patterns are far steadier.

While the vertical motions in supercells are enhanced by the vertical pressure forces, in the mechanism reported in this study the dynamic vertical pressure gradient will tend to oppose the peak downdraft. First, recall that the vortex results from stretching of a column of air with vorticity through strong downward acceleration near the bottom of the column. This strong downward acceleration in the lower portions of the column is associated with the development of the downburst below cloud base. For a column with uniform vorticity, this stretching will enhance the vorticity within the column with the maximum vorticity production located above the strongest downdraft where the vertical gradient $w$ is large. Since we have already shown that when the rotation lies above the peak downdraft the vertical pressure force associated with rotation will tend to oppose the downward acceleration, this illustrates that the rotation produced by this mechanism will generally oppose the magnitude of the strongest downdraft.

7. Summary and discussion

In this study we simulated the life cycle of a rotating downburst similar to the observed cells reported in Kessinger et al. (1988). The rotating downburst was produced within a nonsteady convective system. The rotation was generated by a mechanism that depends on the nonsteadiness of the storm, as vorticity was first generated at middle levels within the updraft and then, as the updraft collapsed and a downdraft formed, the vorticity was subsequently stretched by the downdraft. The mechanism reported herein relies on the vertical shear in the middle troposphere to develop a correlation between the location where vorticity was previously

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Fig. 23. A horizontal cross section of relative flow at 2.0 km at 28.50 s into the simulation. Downdraft magnitudes stronger than $-7.5 \text{ m s}^{-1}$ are shaded. All simulations have constant wind below 3 km and above 9 km. (a) Constant wind in $v$ and linear shear in $u$ varying from a wind 0.0 m s$^{-1}$ at 3 km to 9 m s$^{-1}$ at 7 km. (b) Half-circle veering shear between 3 and 7 km with a radius of 12 m s$^{-1}$. (c) Half-circle backing shear between 3 and 7 km with a radius of 12 m s$^{-1}$.
created by the updraft and the location where the downdraft forms. For a vertical shear vector that veers with height at midlevels, the downdraft will rotate cyclonically. In general, the rotation in downbursts is not likely to increase the magnitude of the downdraft, but tends to slightly retard the downburst magnitude.

One could hypothesize that other mechanisms could produce rotating downbursts in the atmosphere, such as the ambient vertical vorticity being concentrated by convergence into the downdraft. While we concede this possibility, past experience with rotating convective systems suggests that the time scale of this process is generally longer than mechanisms that tilt the ambient vertical shear, since the ambient vertical vorticity is generally relatively small. Another possible mechanism is that proposed for rotating up- and downdrafts in supercells. We have discounted the supercell mechanism for the simulated downburst as the supercell mechanism predicts rotation in the opposite sense to that found in association with the downburst. We also propose that the supercell mechanism is generally not applicable to most downbursts since the significant dynamic characteristic of a supercell is that it is long-lived and, in the limiting case, steady with time, in contrast to the short-lived impulsive circulation commonly associated with most downdraft cases reported in the literature. In addition, supercells are relatively rare, so that the mechanism reported in the study for ordinary cells is likely to be far more common in producing rotation centers than the mechanism proposed for supercells.

The knowledge of how and when a downburst rotates gained from this study is also of significance to nowcasting downburst activity, since the rotation is often easily detectable by Doppler radar. For example, Roberts and Wilson (1989) proposed that the rotation could be detected by Doppler radar prior to the downburst impacting the surface and indicated that the downburst may sometimes be driven by vertical pressure gradients. Although our study demonstrates that rotating downbursts are not likely to be driven by vertical pressure forces, we agree that detecting rotation is relevant to nowcasting downbursts. Although we caution that the maximum rotation will often lag the peak downdraft in time and space, the rotation prior to the downburst reaching the surface will often be sufficiently strong to detect with Doppler radar. One must also keep in mind that the rotation depends on the characteristic of vertical shear profile. In general a strong low-level downdraft will tend to generate cyclonic (anticyclonic) vorticity when the vertical shear vector in the middle levels veers (backs) with height. A vorticity couplet will occur with a strong downdraft and a straight hodograph. Depending upon the magnitude of the vertical shear, the strength of the downdraft, and the wind speed coordinate system of the measurements, the vorticity may be concentrated into a closed vortex in any of these shear profiles. There is also the possibility that the local shear profile that a cloud encounters may be influenced by its neighbors, producing rotational characteristics not expected from the ambient vertical shear profile.

Detection of the rotation associated with the downburst is also of significance to the nowcasting of tornadoes, since the magnitude of the rotation associated with the downdrafts can be large enough so that the forecasting criteria for tornadic activity are satisfied (Kessinger et al. 1988). While the tornado forecasting threshold can be met, however, both the previously mentioned Doppler radar studies and our simulations show that the vorticity at the surface in these rotating downbursts is generally small. Thus, the need to distinguish in real time between downdraft mesocyclones and the parent vortex of a tornadic event is clearly evident if an accurate tornado warning algorithm is to be devised in regions where intense downdrafts often occur.

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REFERENCES


