Hook Echoes and Rear-Flank Downdrafts: A Review

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ABSTRACT

Nearly 50 years of observations of hook echoes and their associated rear-flank downdrafts (RFDs) are reviewed. Relevant theoretical and numerical simulation results also are discussed. For over 20 years, the hook echo and RFD have been hypothesized to be critical in the tornadogenesis process. Yet direct observations within hook echoes and RFDs have been relatively scarce. Furthermore, the role of the hook echo and RFD in tornadogenesis remains poorly understood. Despite many strong similarities between simulated and observed storms, some possibly important observations within hook echoes and RFDs have not been reproduced in three-dimensional numerical models.

1. Introduction

Perhaps the best-recognized radar feature in a horizontal depiction associated with supercells is the extension of low-level echo on the right-rear flank of these storms, called the “hook echo.” Hook echoes are known to be associated with a commonly observed region of subsiding air in supercells, called the “rear-flank downdraft.” Rear-flank downdrafts have been long surmised to be critical in the genesis of significant tornadoes within supercell thunderstorms. In this paper, observations of hook echoes are reviewed first, beginning with the first documentations of the 1950s and 1960s. Rear-flank downdrafts are reviewed next, including discussion of pertinent Doppler radar observations and theoretical and numerical simulation studies. Finally, relatively recent observations made during the Verification of the Origins of Rotation in Tornadoes Experiment (VORTEX; Rasmussen et al. 1994) and smaller subsequent field operations are reviewed. Despite the well-known association among hook echoes, rear-flank downdrafts, and tornadogenesis, their dynamical relationship remains poorly understood. Our current confusion serves as a motivation for the collection of new, in situ observations having unprecedented spatial and temporal detail within hook echoes and rear-flank downdrafts. These data are the foundation of a companion paper (Markowski et al. 2002).

2. Hook echoes

a. Characteristics

The hook echo first was documented by Stout and Huff (1953; Fig. 1) in an Illinois tornado outbreak on 9 April 1953, although van Tassell (1955) is given credit for coining the term. The reflectivity appendage usually is oriented roughly perpendicular to storm motion. Hook echoes are typically downward extensions of the rear side of an elevated reflectivity region (Forbes 1981) called the echo overhang (Browning 1964; Marwitz 1972a; Lemon 1982), with the region beneath the echo overhang termed a weak echo region (Chisholm 1973; Lemon 1977) or vault (Browning and Donaldson 1963; Browning 1964, 1965a). Browning and Donaldson (1963) and Browning (1965b) noted that the southern edge of the hook formed a wall of echo “which was often very sharp and sometimes rather upright.” Hook echoes are typically several kilometers in length and several hundred meters in width, at least as viewed by operational radars (e.g., Garrett and Rockney 1962). A variety of shapes that the hook echo may take were presented by Fujita (1973; Fig. 2).

Fujita (1958a) documented hook echoes associated with other supercells on the same day Stout and Huff made their observations. He inferred the concept of thunderstorm rotation from viewing the evolution of the hook echoes, which he studied in unprecedented (and since unparalleled) detail (Fig. 3). Brooks (1949) earlier had referred to these circulations, having radii of approximately 8–16 km, as tornado cyclones.1 Wind velocity data obtained following the installation of Doppler radars in central Oklahoma in the late 1960s confirmed an association between hook echoes and strong horizontal shear zones associated with storm rotation and tornadoes (e.g., Donaldson 1970; Brown et al. 1973; 1 The tornado cyclone terminology now typically refers to distinct circulations having a scale larger than a tornado but smaller than a mesocyclone (e.g., Agee 1976).
Fig. 1. Radar image from the first documentation of a hook echo. The hook echo was associated with a tornadic supercell near Champaign, IL, on 9 Apr 1953. [From Stout and Huff (1953).]

Lemon et al. 1975; Ray et al. 1975; Ray 1976; Brandes 1977a; Burgess et al. 1977; Lemon 1977; Barnes 1978a,b).

Garrett and Rockney (1962) were the first to relate a circular echo on the tip of a hook echo to the tornado or tornado cyclone. They called this ball-shaped echo an “asc” (annular section of the cylinder of the vortex), but the authors did not offer an explanation for the exact cause of the appearance of the asc. Stout and Huff (1953) also observed a similar feature, but little was discussed of it. Donaldson (1970) noted an echo hole in the tornado he studied, and found that it was collocated with a tornado vortex. Forbes (1981) also observed similar reflectivity features during the tornado outbreak of 3–4 April 1974, as did Fujita and Wakimoto (1982) in their study of the Grand Island, Nebraska, tornadoes of 3 June 1980.

Van Tassell’s (1955) images of a hook echo near Scottsbluff, Nebraska, on 27 June 1955 (it moved directly over the radar) suggested the presence of a faint anticyclonic protrusion from the tip of the hook, extending outward in a direction opposite that of the cyclonic protrusion. An anticyclonic reflectivity flare also has been documented by Brandes (1981), Fujita (1981), and Fujita and Wakimoto (1982). Multiple Doppler radar wind syntheses almost invariably have revealed a region of anticyclonic vorticity on the opposite side of the hook echo as the more prominent (cyclonic) vorticity region (Brandes 1977b, 1978, 1981, 1984a; Ray 1976; Ray et al. 1975, 1981; Heymsfield 1978; Klemp et al. 1981; Fig. 4). It is perhaps surprising that the ubiquity of the vorticity couplet straddling the hook echo largely has been ignored, with the exception of Fujita and his collaborators. Fujita and Wakimoto (1982) documented an anticyclonic tornado within the region of anticyclonic vertical vorticity (Fig. 5). The cyclonic member of the vorticity couplet also was associated with a tornado.

Fujita (1981) proclaimed “Mesoscale modelers should be attracted by such a pair of cyclonic and anticyclonic tornadoes which were evidenced in the Grand Island storm on 3 June 1980 and in the central Iowa storm on 13 June 1977.” However, it was unclear why such attention should be given to the vortex couplet, and the origin of the couplet was not well understood.

b. Formation

Fujita (1958a) originally attributed hook echo formation to the advection of precipitation from the rear of the main echo around the region of rotation associated with the tornado cyclone and updraft. Browning (1964, 1965b) also documented hook echoes and attributed their evolution (Fig. 6) to essentially the same process described by Fujita (1958a). Fujita (1965) later attributed hook echo formation to the Magnus force. He explained that this force pulled the spiraling updraft out of the main echo, resulting in the hook-shaped reflectivity appendage commonly observed on radar displays (Fig. 7).

Fulks (1962) hypothesized that hook echo formation was due to a large convective tower extending into the levels of strong vertical wind shear, which produced cyclonic and anticyclonic flows at opposite ends of the tower—the cyclonic flow to the southwest gave rise to hook echo development. No mention was made of the possibility of an anticyclonic hook echo forming on the north side of the tower from the same mechanism.

Probably no one presented as many detailed Doppler radar analyses of supercells as Brandes [1977a,b, 1978, 1981, 1984a,b; Brandes et al. (1988)]. Brandes (1977a) looked at a nontornadic supercell on 6 June 1974. Hook echo formation was attributed to the “horizontal acceleration of (low-level) droplet-laden air” as the downdrafts intensified and the outflow interacted with the inward-spiraling updraft air. Apparently this hypothesis was essentially that precipitation advection was responsible for hook echo formation, similar to the Fujita (1958a) and Browning (1964, 1965b) hypotheses. A three-dimensional numerical simulation by Klemp et al. (1981) of the Del City, Oklahoma (20 May 1977) supercell also suggested that the horizontal advection of precipitation was important for hook echo development. In some observations of hook echoes associated with tornadoes in nonsupercell storms, the hook echoes also have appeared to result largely from the horizontal advection of hydrometeors (e.g., Carbone 1983; Roberts and Wilson 1995).

Other reliable radar observations have been made that suggest that hook echo formation, in at least some cases, appears to result from the descent of a rain curtain in the rear-flank downdraft (e.g., Forbes 1981; E. N. Rasmussen 2000, personal communication; L. Lemon

1 This was an oral presentation at the VORTEX Symposium in Long Beach, California.
<table>
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<th>Observation/Conclusion</th>
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<td>RFD originated at or above 7 km</td>
<td>Nelson (1977), Lemon et al. (1978), Barnes (1978a), Lemon and Doswell (1979)</td>
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<tr>
<td>RFD originated below 7 km Low ( \theta_v ) at surface in RFD</td>
<td>Klemp et al. (1981)</td>
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<td>Hypothesized the RFD is forced mainly thermodynamically from aloft (which may result from stagnation)</td>
<td>Browning and Ludlam (1962), Browning and Donaldson (1963), Browning (1964), Nelson (1977), Barnes (1978a), Brandes (1981), Klemp et al. (1981)</td>
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<td>Hypothesized the RFD is initiated by dynamic pressure excess aloft but maintained thermodynamically</td>
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<td>Tornadogenesis observed before hook formation</td>
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<td>Tornadogenesis observed at the time of overshooting top collapse Visual observations of clear slots accompanying tornadoes</td>
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<td>Hook echo formation attributed to rotation (but not necessarily the same mechanisms) Hook echoes associated with strong horizontal shears or tornadoes (prior to 1980, in the interest of brevity)</td>
<td>Fujita (1958a, 1965), Fulk (1962), Browning (1964, 1965b), Brandes (1977a)</td>
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<td>Hook echoes located in strong vertical velocity and temperature gradients, somewhat behind the surface windshift associated with the RFD</td>
<td>Marwitz (1972a,b), Burgess et al. (1977), Lemon and Doswell (1979), Brandes (1981)</td>
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<td>Air parcels that enter the tornado pass through the RFD</td>
<td>Brandes (1978), Davies-Jones and Brooks (1993), Wicker and Wilhelmson (1995), Dowell and Bluestein (1997), Adlemman et al. (1999) [and implied by visual observations of Lemon and Doswell (1979), Rasmussen et al. (1982), Jensen et al. (1983)]</td>
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<td>Hypothesized that the RFD is important for tornadogenesis</td>
<td>Ludlam (1963), Fujita (1975b), Burgess et al. (1977), Barnes (1978a), Lemon and Doswell (1979), Brandes (1981), Davies-Jones (1982a,b)</td>
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2000, personal communication). It may be worth the effort in future field experiments to use temporally and spatially high-resolution mobile radar data to systematically investigate the extent to which the hook echo forms from the above process, versus being due to a streamer of precipitation that is extruded from the main echo core by horizontal advection. The relative contribution to hook echo formation from these two processes is not yet known. It is entirely possible that one might dominate in one case, while the other dominates in a different case. Or perhaps the relative contribution from the two processes could be a function of time within a single case.

For completeness, it is noted that at least one documentation has been made of hook echoes not associated with updraft rotation. Houze et al. (1993) showed examples of hook-shaped (in a cyclonic sense, with the hooks pointing toward the right with respect to storm motion) reflectivity structures in left-moving severe storms in Switzerland. These features, termed false hooks by the authors, apparently were associated with the cyclonic downdraft regions on the right (southern) flanks of the anticyclonically rotating storms, in which the updrafts would have been on the left (northern) flanks (Klemp and Wilhelmson 1978a; Wilhelmson and Klemp 1978; Rotunno and Klemp 1982).

c. Tornado forecasting based on hook echo detection

The forecasting potential of hook echo detection began to be explored in the mid-1960s. Sadowski (1958) documented a tornado that occurred after the hook echo became visible (in fact, the hook echo was becoming less discernible on radar at the time of the reported tornado formation). Sadowski might have been the first to speculate that if hook echoes generally preceded tornadogenesis, then it might be possible to issue tornado warnings in advance.

In Stout and Huff’s (1953) report, the evidence had been inconclusive as to whether the hook echo preceded tornadogenesis, or vice versa. In van Tassell’s (1955) summary, it was not mentioned whether the hook echo developed before or after tornadogenesis. The tornado studied by Garrett and Rockney (1962) apparently formed before the hook echo became prominent, unless a narrow hook echo went undetected by the Weather Surveillance Radar-3 (WSR-3) (4° beamwidth) prior to tornadogenesis. The tornado dissipated when the hook “closed off” or merged with the forward-flank echo.

Sadowski (1969) later documented a large amount of success using hook echoes to detect tornadoes within thunderstorms. In a 1953–66 study, he computed an average time of 15 min between hook echo appearance and tornadogenesis in a sample of 13 cases in which hook echoes appeared before tornadoes were reported. Sadowski reported a false alarm rate of only 12%. On the other hand, Freund (1966) found that only 6 of 13 tornadic storms near the National Severe Storms Laboratory in 1964 were associated with hook echoes, and Golden (1974) found that only 10% of waterspouts were associated with hook echoes.

The so-called Super Outbreak of tornadoes on 3–4 April 1974 (Fujita 1975a,b) provided a large sample of a variety of “distinctive echoes” that were studied by Forbes (1975, 1981). Forbes (1975) found that 1) a majority of hook echoes were associated with tornadoes, 2) hook echoes often were associated with tornado families, and 3) tornadoes associated with hook echoes tended to be stronger than those from other echoes. Forbes (1975) also found that, on average, hook echoes appeared 25 min prior to tornadogenesis; however, much variance was present in his sample—10 of 27 (37%) of the hook echoes associated with the first tornado produced by a supercell were detected after the reported1 tornado formation times. Forbes (1981) found a false alarm rate of just 16% when using hook echoes to detect tornadoes. But because hook echoes were relatively rare

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1 The accuracy of the reported tornado times may be questionable for some of the tornadoes studied.
(as he defined them), a less restrictive shape (a “distinctive echo”, e.g., appendages, line-echo wave patterns, etc.) also was considered. Distinctive echoes were associated with a probability of detection of tornadoes of 65%. Forbes (1975, 1981) did raise concern about the generality of his findings, since his statistics were based on the events of a single day. Also, the statistics were based on Weather Surveillance Radar-1957 (WSR-57) data, which may not have adequately resolved finescale echo structures.

3. Rear-flank downdrafts

a. Association with hook echoes

Rear-flank downdrafts (RFDs) are regions of subsid ing air that develop on the rear side of the main updraft of supercell storms, and these regions of descent have a well-established association with hook echoes. The first documentation of an RFD, although not recognized as such, probably was by van Tassell (1955). In that case study and in another by Beebe (1959) on the same storm complex, three “reliable” reports of severe downdrafts on the south side of the Scottsbluff tornado (27 June 1955) were made.

Browning and Ludlam (1962) and Browning and Donaldson (1963) also were among the first to mention the presence of a downdraft in the vicinity of the strongest low-level rotation, behind the main storm updraft. Browning and Donaldson (1963) noted that the hook echo itself may be associated with this downdraft region. Haglund (1969), Fujita (1973, 1979), Lemon et al. (1975), Burgess et al. (1977), Brandes (1977a), Lemon (1977), and Forbes (1981) also documented an association between hook echoes and downdrafts. According to Forbes (1981), “the hook represents a band of precipitation accompanied by downdraft and outflow, surrounding a weak echo region (a region of inflow and updraft).” Brandes (1977a) tentatively concluded that the hook echo reflected downdraft intensification. Haglund (1969) concluded that the hook echo slightly trails the surface wind shift associated with the outflow of the RFD, and that the hook echo is located near the boundary between updraft and downdraft. Surface analyses and aircraft penetrations have revealed that the hook echo is located in a region of large vertical velocity and temperature gradients (Burgess et al. 1977; Marwitz 1972a,b).

b. Visual characteristics

The number of visual and surface observations of supercells increased during the 1970s, largely because of organized storm intercept programs at the National Severe Storms Laboratory (Golden and Morgan 1972; Davies-Jones 1986; Bluestein and Golden 1993). Many of these observations have advanced our understanding of the basic structures associated with tornadoes and their parent storms.

Golden and Purcell (1978) photogrammetrically documented subsiding air on the south side of the Union City, Oklahoma, tornado (24 May 1973), apparently a visual manifestation of the RFD and also
evidence that the tornado occurred in a strong vertical velocity gradient. Moreover, a clear slot was seen to wrap itself at least two-thirds of the way around the tornado. Other observations of clear slots, which are probably always visual manifestations of subsiding air in an RFD,\(^4\) have been presented by Beebe (1959; this was probably the first documentation), Moller et al. (1974), Peterson (1976), Stanford (1977), Burgess et al. (1977), Lemon and Doswell (1979), Marshall and Rasmussen (1982), Rasmussen et al. (1982), and Jensen et al. (1983) (Fig. 8).

Burgess et al. (1977) found that the clear slot could be associated with a hook echo: “Perhaps large droplets are present in the downdraft and are brought down from the echo overhang, even though the air contains only ragged clouds or is visibly cloudless at low levels. If so, since radar reflectivity is more strongly dependent on the size rather than on the number of droplets, radar may show substantial echo in the ‘clear’ slot.” Analysis of the 2 June 1995 Dimmit, Texas, tornadic supercell also indicated an association between the hook echo and clear slot, based on photogrammetrically determined cloud positions (E. N. Rasmussen 2000, personal communication).\(^5\)

c. Surface characteristics

Direct observations within RFDs have been scarce. Tepper and Eggert (1956) appear to have been the first to systematically analyze traces of thermodynamic data near tornadoes, and consequently, within some RFDs. Data were obtained within 25 km of tornadoes in more than 50 cases. Many of the thermograph traces measured only minor fluctuations during the passage of the tornadoes and associated RFDs, and other traces revealed cooling and moistening near the tornadoes. Only a few observations were available within 5 km of the tornadoes, however.

Fujita (1958b) inferred the presence of a surface high pressure annulus encircling the Fargo, North Dakota, tornado cyclone (20 June 1957) from pressure traces in the vicinity of the tornadoes (Fig. 9). Although Fujita speculated that the high pressure was associated with a ring of subsiding air around the tornado, he was unable to verify this speculation. [Ward (1964, 1972) and Snow et al. (1980) found high pressure rings surrounding laboratory and numerically simulated vortices, but it is not clear whether these are the same phenomena inferred by Fujita, which appeared to be of a slightly larger scale.] Surface pressure excesses within RFDs of up to a few millibars also have been documented by subsequent investigators (e.g., Charba and Sasaki 1971; Lemon 1976a; Bluestein 1983), although no one else has documented a high pressure ring as Fujita did.

The studies of the Scottsbluff tornado by van Tassell (1955) and Beebe (1959) contain some of the first descriptions, albeit qualitative, of surface temperature within an RFD at close range from a tornado. At least a couple of observers, located a few hundred meters south of the tornado, reported that the downdrafts felt “cold.” Browning and Ludlam (1962) and Browning and Donaldson (1963) also reported cold temperature

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\(^4\) The absence of a clear slot does not necessarily indicate an absence of subsiding air.

\(^5\) This was an oral presentation at the VORTEX Symposium in Long Beach, California.
and equivalent potential temperature ($\theta_e$) measurements (with respect to the inflow) in the wakes of the Wokingham, England (9 July 1959) and Geary, Oklahoma (4 May 1961) supercells, presumably within the RFDs. Ward (1961) observed "cooler northwest winds a couple miles southwest of the (Geary) tornado."

In a supercell (25 May 1974) investigated by Nelson (1977), the lowest wet-bulb potential temperature ($\theta_w$) values observed at the surface were within the RFD, where they were $\sim$6 K lower than the ambient $\theta_e$ values. Complete separation of the forward-flank downdraft (FFD) and RFD was evidenced by separate temperature minima, as Lemon (1974) earlier had found in another case. Additional observations of relatively low $\theta_e$ and low-$\theta_w$ air at the surface within RFDs have been presented by Brandes (1977a; supercell on 6 June 1974; $\sim$3 K $\theta_e$ decrease), Barnes (1978a,b; supercell on 29 April 1970; 2–3 K $\theta_e$ decrease directly behind the RFD.

In contrast to the findings summarized above, Garrett and Rockney (1962) reported that a warm downdraft was observed about 12–15 km south of a tornado near Topeka, Kansas, on 19 May 1960. The observer described the air as "suddenly becoming noticeably hot, similar to a blast of heat from a stove." Williams (1963) showed that RFD air can arrive at the surface warmer than the surrounding air. He noted that when such an event occurs, it may be south of the hook echo or wherever forced descent is less likely to encounter sufficient liquid water to maintain negative buoyancy.

Fujita et al. (1977) documented a warm RFD (only temperature data were available) near an F4 tornado in the Chicago, Illinois, area on 13 June 1976. On the same day, Brown and Knupp (1980) found nearly constant $\theta_w$ 3 km east of the Jordan, Iowa, F3 and F5 tornado pair, and those observations probably were in the RFD air mass, based on the pressure trace, which measured a pressure excess of a few millibars.

In his summary of Tornado Total Observatory (TOTO) observations, Bluestein (1983) documented a relatively warm RFD and pressure rise of $\approx$2 mb in a nontornadic supercell on 17 May 1981. Bluestein also presented evidence of a 1.5-K temperature rise in an RFD approximately 1.3 km south of the Cordell, Oklahoma, tornado on 22 May 1981. Similar to what Fujita (1958b) first inferred, Bluestein also showed data that suggested high pressure at least partly encircling a tornado (his Fig. 7). In the violent Binger, Oklahoma, tornado on 22 May 1981, only small temperature fluc-
tations (<1 K) were observed as the tornado passed within a few hundred meters north of TOTO (Fig. 10).

Klemp et al. (1981) referred to “cold downdraft (RFD) outflow” in the Del City supercell, but no evidence was presented demonstrating that this air actually was cold—retrieved temperatures by Brandes (1984a) and observations by Johnson et al. (1987) suggested that at least parts of the Del City storm’s rear-flank outflow were warm, although \( \theta_e \) values may not have been as large as in the inflow.

Although there have been surface observations of warm, high-\( \theta_e \) air within RFDs, three-dimensional numerical simulations of supercells almost invariably have produced cold, low-\( \theta_e \) RFDs at the surface (e.g., Klemp et al. 1981; Rotunno and Klemp 1985; Fig. 11). The pioneering numerical modeling studies of supercells conducted in the 1970s (e.g., Schlesinger 1975; Klemp and Wilhelmson 1978a,b; Wilhelmson and Klemp 1978) and the parameter space studies of the 1980s (e.g., Weisman and Klemp 1982, 1984), with no known exceptions, used warm rain microphysics. The relatively simple parameterization was not computationally demanding; thus, experiments requiring large numbers of simulations were feasible. However, the exclusion of ice may have promoted unrealistically excessive amounts of latent cooling near updrafts. When ice physics is included, hydrometeors are distributed over a larger horizontal region, and the intensity of outflow in close proximity to the updraft is reduced (Gilmore and Wicker 1998; Rasmussen and Straka 1998). In some recent, unpublished simulations, relatively warm downdrafts have been produced at the surface when ice physics and a relatively fine spatial resolution (<250 m in the horizontal directions) were used (M. Gilmore 2000, personal communication).

d. Characteristics above the surface

Johnson et al. (1987) presented observations of the RFD and FFD of the Del City storm collected by a 444-m tower as the storm passed overhead. The RFD was associated with \( \theta_e \) values approximately 4 K lower than the ambient conditions; however, the temperature increased 1.5 K and the dewpoint temperature decreased 2.5 K—if \( \theta_e \) was nearly conserved, then air had subsided from approximately 1 km (all heights are above ground level). Although the RFD was not sampled well by the tower, the data that were available suggested the presence of a downward-directed perturbation pressure gradient within the lowest half kilometer.

Dowell and Bluestein (1997) documented the passage of another tornadic supercell over the same instrumented...
Fig. 10. Pressure, wind speed, wind direction, and temperature traces from TOTO near the Binger, OK, tornado on 22 May 1981 from approximately 1922–1931 CST. Only relative wind direction is known; veering (backing) is indicated by a downward (upward) change. The tornado passed approximately 600 m north of TOTO. Note the small (less than $1$ K) temperature fluctuations recorded. It is inferred that the hook echo and RFD associated with this violent tornado were relatively warm. [From Bluestein (1983).]

tower on 17 May 1981. They found large $\theta_e$ and $\theta_v$ (virtual potential temperature) deficits in the portion of the RFD sampled by the tower ($>12$ and $>5$ K, respectively), which was within a region of fairly high ($>40$ dBZ) radar reflectivity west of the circulation center. The deficits were prominent over the entire height of the tower.

With the exception of the direct observations presented by Johnson et al. (1987) and Dowell and Bluestein (1997), thermodynamic quantities above the ground within RFDs have been obtained only by indirect means. Brandes (1984a) and Hane and Ray (1985) were among the first to use the methods proposed by Gal-Chen (1978) and Hane et al. (1981) to retrieve thermodynamic fields in supercells from three-dimensional wind fields synthesized from multiple Doppler radars. Brandes (1984a) retrieved the pressure and buoyancy fields in the Del City and Harrah, Oklahoma (8 June 1974) tornadic storms. At 3.3 km on the rear side of the updraft of the Del City storm, relatively cold temperatures were retrieved within the RFD—radar reflectivity was a minimum here, possibly implying that evaporation was occurring. Behind the rear-flank gust front in the eastern mesocyclone quadrants, evidence was found of warm temperatures at low levels during the tornadic stage. The relatively warm conditions were attributed to subsidence within the RFD. In the Harrah storm, Brandes (1984a) also retrieved negative buoyancy in the RFD aloft, but no mention was made of the low-level buoyancy immediately behind the gust front in the eastern quadrants of the mesocyclone.

Hane and Ray (1985) also completed a thermodynamic retrieval for the Del City storm. In the pre-tornadic stage, the pressure distribution included at each level a high–low couplet across the updraft with the maximum horizontal pressure gradient generally oriented along the environmental shear vector at that altitude, in agreement with linear theory predictions (Rotunno and Klemp 1982). Although the orientation of the horizontal pressure gradient agreed relatively well with linear theory, its magnitude did not agree as well. The authors stated that possibly the orientation and magnitude of the environmental shear vector were not known exactly (also, calculations of the horizontal vertical velocity gradient may have had significant errors). In the tornadic stage, the pressure field contained a pronounced minimum at low levels coincident with the mesocyclone, probably due to strong low-level vertical vorticity. Hane and Ray found weak high perturbation pressure ($\sim 1$ mb) in the RFD at low levels behind the gust front during the time of the tornado. Vertical gradients of nonhydrostatic pressure perturbations may be relevant to the formation and evolution of the RFD, as will be discussed in the next subsection. The RFD contained significant negative buoyancy at low levels in Hane and Ray’s analysis (temperature deficits as low as $-4.5$ K; Fig. 12).

The retrieval results of Brandes (1984a) and Hane and Ray (1985) in the Del City storm were in relatively close agreement with the direct measurements within the inflow and FFD regions reported by Johnson et al. (1987). But Johnson et al. cautioned that “noticeable differences in the RFD region suggested that there was room for improvement in the retrieval methods.” Brandes and Hane and Ray had used considerable smoothing on the buoyancy field to eliminate noise; the details in the retrieved low-level buoyancy fields may have been suspect.

e. Origins and formation

In reviewing previous studies and hypotheses pertaining to RFD formation, it may be helpful to begin

7 The RFD had wrapped cyclonically around the updraft, so that downdraft air also was found in the eastern (forward) quadrants of the mesocyclone. This “occlusion process” will be discussed in section 4.
Fig. 11. Contours at $z = 250$ m of (a) potential temperature fluctuation ($\bar{\theta} = 303$ K), in intervals of 1 K; and (b) equivalent potential temperature fluctuation ($\bar{\theta}_e = 340$ K), in intervals of 2 K, in the simulation conducted by Rotunno and Klemp (1985). The updraft is represented by the shaded region. The thick circular line encloses the region where the cloud water is greater than 0.1 g kg$^{-1}$ (wall cloud). The thick dashed line encloses the region where the cloud water is greater than 0.1 g kg$^{-1}$ at $z = 500$ m. [Adapted from Rotunno and Klemp (1985).]

Fig. 12. Horizontal distribution of buoyancy and vector horizontal wind at 1 km at 1847 CST 20 May 1977 in the Del City, OK, storm. Buoyancy (potential temperature fluctuation) has been filtered to remove noise, and is contoured at 1.5°C intervals (negative values dashed). Updraft maxima are denoted by $\bigodot$ and vorticity maxima are denoted by $\bigotimes$. [From Hane and Ray (1985).]

with an inspection of the inviscid vertical momentum equation written as

$$\frac{d\bar{\theta}}{dt} = \bar{\theta} + \bar{\theta} \cdot \nabla \bar{w} = -c_p \frac{\partial \pi}{\partial z} + B, \quad (1)$$

where $d\bar{\theta}/dt$ is the vertical acceleration following a parcel, $\bar{\theta} \cdot \nabla \bar{w}$ is the local vertical acceleration, $\bar{w}$ is the three-dimensional velocity vector, $-c_p \pi$ is the advection of vertical velocity, $\bar{\theta}$ is the mean potential temperature, $\pi$ is the perturbation Exner function, and $B$ is the buoyancy which can be written as

$$B = g \left( \frac{\bar{\theta}'}{\bar{\theta}} + 0.61 q'_w - q_l - q_i \right), \quad (2)$$

where $\bar{\theta}'$ and $q'_w$ are potential temperature and water vapor mixing ratio fluctuations from the base state, respectively, and $q_l$ and $q_i$ are liquid water (includes cloud water and rain water) and ice mixing ratios, respectively.

By taking the divergence of (1) and assuming that $\nabla \pi \sim -\pi$ (a reasonable assumption for “well-behaved” fields away from the boundaries), it can be shown (Rotunno and Klemp 1982) that
\[ \pi = \left[ \left( \frac{\partial u'}{\partial x} \right)^2 + \left( \frac{\partial v'}{\partial y} \right)^2 + \left( \frac{\partial w'}{\partial z} \right)^2 \right] + \frac{1}{2} \left| D \right|^2 - \left| \eta \right|^2 \]

where \( |D| \) and \( |\eta| \) are the magnitudes of the total deformation and vorticity, respectively. The \( (\partial u/\partial x)^2 + (\partial u/\partial y)^2 + (\partial w/\partial z)^2) \) terms refer to the fluid extension terms. If (3) is linearized about a base state containing vertical wind shear (primed velocity components represent fluctuations from the base state, which is given by \( \nabla(z) = [\nabla(z), \nabla(z), 0] \)), it may be rewritten as

\[ \pi = \left( \frac{\partial u'}{\partial x} \right)^2 + \left( \frac{\partial v'}{\partial y} \right)^2 + \left( \frac{\partial w'}{\partial z} \right)^2 + 2 \left( \frac{\partial u'}{\partial x} \frac{\partial v'}{\partial y} + \frac{\partial v'}{\partial x} \frac{\partial u'}{\partial y} + \frac{\partial w'}{\partial x} \frac{\partial u'}{\partial z} + \frac{\partial w'}{\partial y} \frac{\partial v'}{\partial z} \right) \]

\[ + 2 \frac{\partial \mathbf{V}}{\partial z} \cdot \nabla w' - \frac{\partial B}{\partial z} \]

\[ = \pi_{nl} + \pi_r + \pi_b \]

\[ = \pi_{nl} + \pi_r + \pi_b \]

where \( \pi_{nl}, \pi_r, \) and \( \pi_b \) are the contributions to \( \pi \) from nonlinear, linear, and buoyancy effects, respectively.

\[ \pi_{nl} \propto \left[ \left( \frac{\partial u'}{\partial x} \right)^2 + \left( \frac{\partial v'}{\partial y} \right)^2 + \left( \frac{\partial w'}{\partial z} \right)^2 \right] \]

\[ + 2 \left( \frac{\partial u'}{\partial x} \frac{\partial v'}{\partial y} + \frac{\partial v'}{\partial x} \frac{\partial u'}{\partial y} + \frac{\partial w'}{\partial x} \frac{\partial u'}{\partial z} + \frac{\partial w'}{\partial y} \frac{\partial v'}{\partial z} \right) \]

\[ \pi_r \propto \frac{\partial \mathbf{V}}{\partial z} \cdot \nabla w' \]

\[ \pi_b \propto -\frac{\partial B}{\partial z} \]

and \( \pi_{nl} \) collectively refers to the nonlinear and linear effects as “dynamic” effects on \( \pi \). Equation (7) indicates that nonlinear dynamic high (low) pressure perturbations are associated with convergence and divergence and deformation (rotation). Equation (8) indicates that linear dynamic high (low) pressure perturbations are located upshear (downshear) of an updraft. Equation (9) indicates that high (low) pressure perturbations due to buoyancy are located above (below) the level of maximum buoyancy.

Using (5) and (6), the vertical momentum equation can be written as

\[ \frac{\partial w}{\partial t} + \mathbf{v} \cdot \nabla w = -c_p \frac{\partial \pi_{nl}}{\partial z} - c_p \frac{\partial \pi_r}{\partial z} \]

\[ + \left( -c_p \frac{\partial \pi_b}{\partial z} + B \right) \]

\[ = -c_p \frac{\partial \pi_{nl}}{\partial z} + \left( -c_p \frac{\partial \pi_b}{\partial z} + B \right) \]

where \( -c_p \frac{\partial \pi_{nl}}{\partial z} \) sometimes is referred to as the dynamic forcing and \( -c_p \frac{\partial \pi_b}{\partial z} + B \) sometimes is referred to as the buoyancy forcing. If the vertical gradients of the fluid extension terms are neglected, along with the vertical gradients of the deformation and horizontal vorticity, then it can be shown that

\[ \frac{\partial \pi_{nl}}{\partial z} \approx -\frac{\partial^2 \zeta}{\partial z^2} \]

\[ \frac{\partial \pi_r}{\partial z} \approx -\frac{\partial}{\partial z} \left( \frac{\partial \mathbf{V}}{\partial z} \cdot \nabla w \right) \]

\[ \frac{\partial \pi_b}{\partial z} \approx -\frac{\partial B}{\partial z} \]

where \( \zeta \) is the vertical vorticity.

From (11) it is evident that descent can arise owing to negative buoyancy, which can be generated according to (2) by cold anomalies produced by evaporative cooling or hail melting, or by precipitation loading, and by vertical perturbation pressure gradients that can arise from, according to (12)–(14), vertical gradients of vertical vorticity, “stagnation” of environmental flow at an updraft, and pressure perturbations due to vertical buoyancy variations (which are partially due to hydrostatic effects), respectively. Research presented in the past 40 years has found that all of the terms in (11) can be significant.

Browning and Ludlam (1962) and Browning and Donaldson (1963) suggested that the RFDs in the Wokingham and Geary supercells might have been driven by negative buoyancy (i.e., “thermodynamically” forced) due to evaporation. Browning (1964) surmised that the rightward propagation of supercells increased

\[ \text{Note that the use of the term stagnation here does not imply that supercell updrafts are solid obstacles, as has been suggested by several investigators in the past (e.g., Newton and Newton 1959; Fujita 1965; Alberty 1969; Charba and Sasaki 1971; Brown 1992). Theoretical studies have exposed serious weaknesses in the obstacle analogy (e.g., Rotunno 1981; Rotunno and Klemp 1982; Davies-Jones et al. 1994), and these studies also have shown that the pressure distribution around an updraft is not what would be expected if the updraft was behaving as an obstacle, except at the storm top (Davies-Jones 1985). Some studies have shown that updrafts occasionally can display behavior that appears similar to how a solid obstacle might be expected to behave (e.g., Lemon 1976b; Klemp et al. 1981).}

\[ \text{For additional discussion of downdraft forcings in terms of the vertical momentum equation, the reader is referred to the review by Knupp and Cotton (1985).} \]
their midlevel storm-relative flow so as to increase evaporative cooling, and ultimately aid in the genesis of downdrafts (both on the rear and forward storm flanks). These hypotheses were proposed at least partly because of findings by Browning and Ludlam (1962) and Browning and Donaldson (1963) of low $\theta_u$ air in the wakes of the Wokingham and Geary storms, which apparently had midlevel origins.

Brandes (1981) also concluded that RFDs are initiated by the production of negative buoyancy aloft: “presumably the initiating downdraft (associated with the rear-flank gust front) is formed by precipitation falling from the sloping updraft . . . we suppose the intruding flow has low $\theta_u$, and when chilled by evaporation, becomes negatively buoyant . . . because the entrained air penetrated well into the storm, evaporative cooling rather than perturbation pressure forces may initiate the downdraft.” Brandes (1984a) made a similar claim, based on retrieved buoyancy: “at 3.3 km, cool temperatures on the southern fringe of the storm were suggestive of evaporative cooling as environmental air mixed with storm air.”

Klemp et al. (1981) attributed the RFD in the Del City storm to water loading and evaporation based on precipitation trajectories crudely approximated using estimated terminal fall speeds. Moreover, midlevel flow approaching the storm from the east flowed through the FFD—not through the RFD as Browning (1964) had conceptualized. RFD air at the surface appeared to have come from 1–2 km above the ground, directly behind the gust front, based on trajectory analyses in their numerical simulation and observations of the storm. Air from higher levels reached the surface further behind the storm.

Nelson (1977) found an erosion of the hydrometeor field at and below 7 km, as well as a sharp reflectivity gradient on the west flank of an Oklahoma multicell storm that evolved into a supercell on 25 May 1974—these radar observations were believed to have been a manifestation of RFD formation that apparently occurred at the start of the transition from multicell to supercell. Nelson noted two mechanisms suggestive of RFD formation—evaporative cooling and/or dynamic pressure perturbations (presumably he was referring to those related to linear effects, i.e., stagnation). Nelson believed that the evaporation-driven effect was more likely because of the echo erosion aloft; he also cited strong storm-relative winds (~16 m s$^{-1}$ in the 7–9 km layer) and a large dewpoint depression (~21 K) at the level of apparent RFD formation. Forbes (1981) found similar radar signatures suggesting echo erosion and RFD formation during the 3–4 April 1974 tornado outbreak.

Lemon et al. (1978) and Barnes (1978a) also concluded that the RFD forms at middle to upper levels. Lemon et al. based their findings on an analysis of the Union City tornadic supercell. They analyzed a persistent diffluent flow region in the 7–10 km layer northwest (upshear) of the mesocyclone that they believed was associated with a downdraft. Barnes’ conclusion that the RFD formed between 6.0–7.5 km was based on his study of tornadic storms in Oklahoma on 29–30 April 1970. He surmised that the storm-relative midlevel flow (20–25 m s$^{-1}$) approaching the cyclonically rotating updraft was decelerated and deflected on the upwind (south) side while the relative upwind stagnation point shifted to the left of the intercepting wind vector; that is, toward the southwest flank. Here “stagnating” air experienced the longest contact with the adjacent updraft while mixing only slightly with it—both cloud and small precipitation drops chilled this air by evaporation and began its downward acceleration before saturation could occur. Barnes added “We emphasize that the high horizontal momentum and proximity to the updraft make the RFD a potentially important interactant with the gust front and updraft’s surface roots . . . We also note that the location and extent of such a downdraft probably depends upon the ambient flow relative to the storm, which very likely requires a specific vertical shear profile to place it on the rear flank of a storm where it attains an influential position.” Barnes interpreted the large reflectivity gradient on the midlevel upwind (southwest) flank as indicating dry ambient air adjacent to a precipitation-laden updraft. Bonesteele and Lin (1978) made a similar inference.

Lemon and Doswell (1979) developed a conceptual model of a supercell from an extensive compilation of surface, visual, and radar observations (Fig. 13). This model included an FFD and RFD, a surface gust front structure resembling a midlatitude cyclone, a hook-shaped reflectivity region surrounding a cyclonically rotating updraft, and a tornado, if present, that resided within the vertical velocity gradient between the updraft and RFD. This model has undergone little modification since its presentation over 20 years ago. Based largely on the work by Barnes (1978a,b), Lemon and Doswell inferred that the RFD typically originated between 7–10 km on the relative upwind side of the updraft [note that they did not say upshear side; refer to (8); Rotunno and Klemp (1982) showed that the linear forcing for pressure fluctuations depends on the vertical shear, and numerical results confirmed this theoretical prediction, as did some later dual-Doppler radar findings (e.g., Hane and Ray 1985)]. The authors cited the observation of an echo-free hole at 7.5 km, directly above a notch behind the low-level hook echo—they believed this to be the signature of the RFD. Lemon and Doswell proposed that storm-relative inflow impingement was the RFD source, because Darkow and McCann (1977) showed that the relative flow at these levels is much stronger than the storm-relative flow minimum they found at 4 km, and because of the Barnes (1978a,b) and Nelson (1977) observations. Lemon and Doswell also hypothesized that the RFD initially is dynamically forced, and then enhanced and maintained by precipitation drag and evaporative cooling.
Brandes (1984a) retrieved excess pressure aloft on the rear of the Del City storm, which may have suggested that the RFD was partly forced by a downward-directed dynamic pressure gradient, as Lemon and Doswell had proposed, but the vertical pressure gradients in the stagnation region could not be examined due to a paucity of scatterers and corresponding lack of radar velocity data. (A lack of data due to the pristine air common in RFD regions probably still is one of the biggest obstacles in understanding the formation mechanisms and the role of the RFD today.)

Klemp et al. (1981) simulated a supercell with a composite sounding derived from three “proximity” soundings on 20 May 1977, and compared the simulated storm characteristics to those observed in the Del City storm. “Trajectory” analysis (these were not true trajectories, but rather streamlines—if the storm was assumed to be quasi-steady, then the streamlines would be similar to trajectories) in the simulated supercell showed obstacle-like flow at 7–10 km. Parcels at 7 km that impinged upon the upshear side of the updraft did not appear to sink [in contrast to observations made in different cases by Barnes (1978a), Nelson (1977), and Lemon and Doswell (1979)], but those at 4 km did; that is, the RFD apparently was 4–7 km deep (parcels from 4–7 km did not reach the surface, but negative vertical velocities extended to 4–7 km). Directly behind the gust front, RFD air appeared to come from 1–2 km aloft; farther behind the gust front, air from higher levels reached the surface.

For the sake of completeness, it might be worth mentioning that Shapiro and Markowski (1999) recently investigated the formation of downdrafts in simple two-layer vortices using an analytic model. The applicability of the idealized model to real atmospheric vortices, in which buoyancy, buoyancy gradients, precipitation, and asymmetries probably are important, is questionable. Their results demonstrated how the “vortex valve” effect (Lemon et al. 1975; Davies-Jones 1986) can transport vorticity from the top of a homogeneous, axisymmetric, rotating fluid to low levels via an annular downdraft and secondary circulation, when the top layer of fluid rotates with an angular velocity larger than that of the bottom layer of fluid.

Prior to 1983, investigators sought forcing for the RFD from middle and upper levels, as has been reviewed in this section. But downdraft forcing also can arise at low levels. This is the subject of the next section.

4. Occlusion downdrafts

a. Evolution as simulated in numerical models

Klemp and Rotunno (1983) investigated the transition of a supercell into its tornadic phase through use of a high-resolution (250-m horizontal grid spacing) model initiated within the interior of the domain of the Del City supercell simulation performed by Klemp et al. (1981). With the enhanced resolution, Klemp and Rotunno found that the low-level cyclonic vorticity increased dramatically and the gust front rapidly occluded as small-scale downdrafts developed in the vicinity of the low-level circulation center. They concluded that the intensification of the RFD during the occlusion process was dynamically driven by the strong low-level circulation, that is, by way of a dynamic perturbation pressure gradient such as that given by (12). This was the first study to propose such a mechanism for downdraft genesis and intensification. Later, Brandes (1984a,b), Hane and Ray (1985), and Brandes et al. (1988; see section 3) made the same conclusion based on Doppler radar analyses of tornadic storms, as did Trapp and Fiedler (1995), Wicker and Wilhelmson (1995), and Adlerman et al. (1999) based on numerical simulations.

Klemp and Rotunno (1983) defined the RFD as the downdraft “which supports the storm outflow behind the convergence line on the right flank.” They stated that since nontornadic storms often were observed to persist for long periods of time with a well-defined gust front, these storm-scale downdrafts were not uniquely linked to tornadogenesis within a storm. On the other hand, noted Klemp and Rotunno, if a storm progressed into a tornadic phase, the gust front became occluded

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10 Idealized three-layer vortices also were investigated using a numerical model.
Fig. 14. Low-level flow field \((z = 1 \text{ km})\) in Klemp and Rotunno’s (1983) nested, 250-m horizontal resolution simulation of the Del City, OK, supercell (20 May 1977) at the time that an occlusion downdraft was observed. The region in which the cloud water exceeds 0.4 g kg\(^{-1}\) is shaded. The heavy black line encloses the region where the rain water mixing ratio exceeds 0.5 g kg\(^{-1}\). The circulation center is indicated with the black dot. Vertical velocity is contoured at 5 m s\(^{-1}\) intervals. What has been referred to as the occlusion downdraft is located within a broader region of negative vertical velocities, which has been referred to as the rear-flank downdraft. [Adapted from Klemp and Rotunno (1983).]

The finding of Klemp and Rotunno that the occlusion downdraft is driven by low-level rotation sometimes has been implied as being in conflict (e.g., Carbone 1983; Brandes 1984a; Klemp 1987) with their predecessors’ early hypotheses that the RFD is driven from aloft thermodynamically or dynamically and is responsible for increasing low-level rotation (e.g., Fujita 1975b; Burgess et al. 1977; Barnes 1978a; Lemon and Doswell 1979). However, it is the author’s opinion that the apparent conflict may be one of semantics. If the occlusion downdraft and RFD are to be viewed as two distinct entities, as proposed by Klemp and Rotunno, then the formation mechanisms and roles of the occlusion downdraft and RFD should not be anticipated to be necessarily identical. Observations of RFD formation preceding the increase of vorticity near the ground are plentiful (e.g., Barnes 1978a; Lemon and Doswell 1979; Brandes 1984a,b). Once a downdraft forms, the distribution of vortex lines invariably must be affected according to Helmholtz’s theorem, and feedbacks on the downdraft by the new vorticity distribution would be probable (e.g., when rotation near the ground becomes substantial, a downward-directed vertical pressure gradient could become established, accelerating air toward the ground). Early hypotheses that the RFD is responsible for bringing rotation to low levels never asserted that once low-level rotation began to intensify that dynamic effects could not feed back on the downdraft. Therefore, it is believed that the occlusion downdraft can be viewed as a rapid, small-scale intensification of the RFD that occurs after the RFD transports larger angular momentum air toward the ground. Stated another way, the occlusion downdraft may be viewed as a by-product of the near-ground vorticity increase, and vorticity cannot become large next to the ground without the RFD (Davies-Jones 1982a,b; Davies-Jones and Brooks 1993; Brooks et al. 1993, 1994; Wicker and Wilhelmson 1995); that is, the RFD is ultimately necessary for occlusion downdraft formation. Perhaps not surprisingly, observations have not been made in supercell storms of an occlusion downdraft preceding or occurring in the absence of an RFD, or even at a location not within an RFD.\(^{11}\)

It might be tempting to argue that the RFD and occlusion downdraft should be considered separate down-

\(^{11}\) In some nonsupercell tornadoes in which preexisting vertical vorticity is present at the surface, an RFD is not needed to transport circulation to low levels in order for tornadogenesis to occur. It might be possible for the low-level vorticity amplification associated with this tornadogenesis process to dynamically induce a downdraft similar to that which Klemp and Rotunno (1983) called an occlusion downdraft in a supercell storm; thus, it might be possible for an occlusion downdraft to occur in the absence of an RFD in such a case. Carbone (1983) compared a downdraft he observed in a nonsupercell tornado event to the occlusion downdraft defined by Klemp and Rotunno. However, the dominant forcing for the downdraft was uncertain; thus, it is not known whether the downdraft documented by Carbone (1983) should be regarded as an occlusion downdraft, at least as defined by Klemp and Rotunno.
drafts because the dominant forcings are different. It is the author's opinion that a contiguity criterion should be considered when debating whether two phenomena having different forcings are regarded as "separate." Otherwise, the updraft of a supercell should be viewed as two separate updrafts, with one updraft being driven by nonhydrostatic pressure gradient forces below the level of free convection, and another updraft being driven largely by buoyancy forces above the level of free convection. The evolution put forth in the paragraph above is not at odds with early proposals that the RFD is initiated at middle to upper levels and is responsible for initiating rotation near the ground, nor is it in conflict with contentions that strong subsidence develops near the tornado during or after its formation.

It is speculated that the clear slot may be a visual manifestation of an intensifying RFD or occlusion downdraft. Updrafts also have been shown to weaken during the stage when low-level rotation rapidly increases, probably also due to the formation of a downward-directed dynamic pressure gradient induced by the rotation (e.g., Brandes 1984a,b). Fujita (1973), Lemon and Burgess (1976), and Burgess et al. (1977) have documented the collapse of overshooting storm tops near the time of tornadogenesis, which presumably is a manifestation of updraft weakening.

It also should be noted that although the occlusion downdraft in the Klemp and Rotunno (1983) simulation was found to be driven by low-level vertical vorticity amplification, the occlusion downdraft did not descend along the axis of low-level rotation. An explanation was not offered. One might expect that the vertical pressure gradient associated with the vertical gradient of vertical vorticity would lead to a maximum acceleration along the rotation axis. Two reasons might account for the asymmetry: 1) the dynamic vertical perturbation pressure gradient associated with the vertical gradient of vertical vorticity squared \((\partial \zeta^2/\partial z)\) does not contribute to vertical velocity directly, but rather to vertical accelerations—thus, \(dw/dt\) might be a minimum in the vorticity maximum center, but if this occurs within the updraft (where \(w \gg 0\)), then a downdraft (\(w < 0\)) may first appear on the periphery of the updraft, away from the center of rotation, where \(w\) is less positive; and 2) other terms in the vertical momentum equation, when combined with the dynamic vertical perturbation pressure gradient force, may force the strongest downward acceleration away from the axis of largest vertical vorticity—for example, the buoyancy forcing may favor ascent in the updraft center, so that the net effect of the buoyancy forcing and dynamic vertical perturbation pressure gradient may lead to the strongest downward acceleration on the updraft periphery. Superposition of the fields of the vertical momentum equation forcing terms in Klemp and Rotunno's (1983) simulation leads to the strongest downward acceleration being to the southeast of the maximum low-level rotation (Fig. 15). Thus, it is entirely possible for an occlusion downdraft to be "driven" by low-level rotation even if the occlusion downdraft is not collocated with the low-level rotation.12

b. Observations

Brandes (1978; the Harrah storm) appears to have made observations prior to the Klemp and Rotunno (1983) simulation of a downdraft not becoming prominent until after low-level rotation became substantial. Brandes (1981) also stated, following his analysis of the Del City storm, that "the sudden appearance of strong rear downdrafts in storms persisting for hours may also relate to the intensity and distribution of updrafts and vorticity." Brandes (1984a) attributed sudden occlusion downdraft formation in the Del City and Harrah storms to the vertical pressure gradient owing to the explosive growth of low-level vorticity as Klemp and Rotunno (1983) found. Furthermore, Brandes’s data also showed that the occlusion downdraft did not descend along the axis of the strong low-level vorticity. Brandes (1984b) claimed that the occlusion downdraft formed after the incipient tornado had been detected, and roughly coincided with tornado formation. Hane and Ray (1985) also documented occlusion downdraft formation in the Del City storm.

Based on their analyses of the Lahoma and Orienta, Oklahoma, supercells (2 May 1979), Brandes et al. (1988) hypothesized that because RFDs possess weak positive or negative helicity (because couplets of vertical vorticity straddle RFDs), the decline of storm circulation might be hastened by turbulent dissipation when the downdraft air eventually mixes into supercell updrafts. As did Brandes (1984a,b) and Klemp and Rotunno (1983), Brandes et al. claimed that "the updraft minimum in the Lahoma storm and RFD in the Orienta storm apparently owed their existence to the build-up of low-level vorticity and related downward vertical pressure gradients." Large downward pressure forces existed within the RFD and left-hand portions of the persistent updraft region in the Orienta storm, and to the rear of the persistent updraft in the Lahoma storm. Brandes et al. (1988) probably presented the most comprehensive analyses, discussion, and insight into the pressure distribution in supercells to date.

5. Role of RFDs in tornadogenesis

a. Observations-based hypotheses

Ludlam (1963) was one of the first to write that downdrafts, especially those located on the rear flank of supercells, actually may be important to tornadogenesis: "It is tempting to look for the spin of the tornado in

12 Wakimoto and Cai (2000) proposed that it also may be possible for an occlusion downdraft to reach the surface away from the center of strongest near-ground vorticity if a mesocyclone is vertically tilted.
the vorticity present in the general air stream as shear and tilted appropriately in the vicinity of the interface between the up- and down-motions.” Fujita (1975b) also proposed that the downdrafts associated with hook echoes may be fundamentally critical to tornado formation, in terms of his “recycling hypothesis”: 1) downdraft air is recirculated into the (developing) tornado, 2) this process results in an appreciable convergence on the back side of the (developing) tornado, and 3) the downward transport of the angular momentum by precipitation and the recycling of air into the tornado will create a tangential acceleration required for the intensification of the tornado. Research conducted with the aid of coherent radars in the ensuing years led others...
(e.g., Burgess et al. 1977; Barnes 1978a; Lemon and Doswell 1979; Brandes 1981) to make the same general speculation. Burgess et al. (1977) believed that the RFD, hook echo, and tornadogenesis were intimately connected: “The formation and evolution of the RFD is judged extremely important to tornado formation. . .the severe tornado [the Oklahoma City tornado of 8 June 1974] appears related to the increased vorticity source provided by presumed downdraft intensification and gust front acceleration along the right flank.” Forbes (1981) also discovered signatures (e.g., a sharp reflectivity gradient along the upshear side of the updraft, and occasionally a small echo mass several kilometers to the right of the right-rear edge of the main echo) suggesting RFD formation (1–10 min) prior to tornadogenesis.

Lemon and Doswell (1979) noted that just before tornadogenesis, the mesocyclone center shifted from near the updraft center to the zone of high vertical velocity gradient. The early mesocyclone apparently was a rotating updraft, whereas the transformed mesocyclone had a divided structure, with strong cyclonically curved updrafts to the east in the “warm inflow sector” and strong cyclonically curved downdrafts to the west in the “cold outflow sector.” And while the tornado was apparently found in a strong vertical velocity gradient, Lemon and Doswell noted that it probably was located on the updraft side of that gradient.

Lemon and Doswell explained the evolution of the RFD and tornadogenesis as follows: 1) air decelerates at the upwind stagnation point, is forced downward, and mixes with air below, which then reaches the surface through evaporative cooling and precipitation drag; 2) the initially rotating updraft is then transformed into a new mesocyclone with a divided structure, in which the circulation center lies along the zone separating the RFD from the updraft (this process appears to result, in part, from tilting of horizontal vorticity); and 3) “descent of the mesocyclone circulation occurs simultaneously (within the limits of temporal resolution) with the descent of the RFD.”

Observations of low-level vorticity couplets within RFDs that seem to straddle the hook echoes (introduced in section 2a) may be indications that tilting of vorticity by the RFD is important in the formation of tornadoes within supercell storms, as hypothesized by Davies-Jones (1982a,b) and Davies-Jones and Brooks (1993). Furthermore, it may be worth noting that during the tornadogenesis phase in supercells, the parcels of air that enter the tornado or incipient tornado regularly seem to pass through the hook echo and RFD (Brandes 1978; Klemp et al. 1981; Dowell and Bluestein 1997; J. Wurman et al. 2000, personal communication; Fig. 16), which may have been the basis for Fujita’s (1975b) recycling hypothesis. Visual observations of mesocyclones being nearly totally occluded by the RFD, as evidenced by observations of the clear slot during and just prior to the tornadic stage, such as those made by Lemon and Doswell (1979), Rasmussen et al. (1982), and Jensen et al. (1983), also may imply that the air entering the tornado comes from the RFD. The possible importance of the clear slot also did not escape the attention of Ludlam (1963): “. . . often the funnel is photographed spectacularly against a segment of bright and practically cloud-free sky beyond the edge of the arch cloud.” In addition to the observational evidence, simulations also have indicated that air parcel trajectories pass through the RFD en route to intensifying near-ground circulations (Davies-Jones and Brooks 1993; Wicker and Wilhelmson 1995; Adlerman et al. 1999).

Davies-Jones (1998) recently has questioned whether the hook echo is really a passive indicator of a tornado, since close-range airborne and mobile radar observations during recent field experiments have revealed hook echo formation prior to tornadogenesis in every case. Davies-Jones hypothesized that the hook echo may actually instigate tornadogenesis, either baroclinically, by way of buoyancy gradients within the hook echo (Davies-Jones 2000a), or barotropically, by redistributing angular momentum (Davies-Jones 2000b).

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13 This was a poster presented at the 20th Severe Local Storms Conference in Orlando, Florida.

14 The “arch cloud” Ludlam refers to was that “on the forward right flank of severe storms, outside of the precipitation region;” that is, that which is on the leading edge of the trailing gust front.

15 Previous studies documenting cases in which tornadoes were reported prior to hook echo detection (e.g., Garrett and Rockney 1962; Forbes 1975) relied on temporally and spatially coarser radar data compared to what was available in field experiments such as VORTEX, and reported tornado times may not always have been accurate to within a few minutes.
b. Theoretical considerations

Most of the theoretical and numerical modeling studies pertaining to supercell storms have investigated the development of midlevel and low-level rotation by way of tilting of horizontal vorticity (either associated with the large-scale mean vertical wind shear or generated solenoidally by a baroclinic zone) by an updraft (e.g., Rotunno 1981; Rotunno and Klemp 1982, 1985; Lilly 1982, 1986a,b; Davies-Jones 1984). However, Davies-Jones (1982a,b) noted that in order to obtain large vertical vorticity at the ground in an environment in which vortex lines are initially quasi-horizontal, a downdraft would be necessary. Tornadoes may arise in the absence of a downdraft in environments containing preexisting vertical vorticity at the surface, such as in some cases of “nonsupercell tornadogenesis” (e.g., Wilson 1986; Wakimoto and Wilson 1989; Roberts and Wilson 1995; Lee and Wilhelmson 1997a,b, 2000).

Davies-Jones (1982a,b) concluded that in a sheared environment with negligible background vertical vorticity, an “in, up, and out” circulation driven by forces primarily aloft would fail to produce vertical vorticity close to the ground [this conclusion depends on eddies being too weak to transport vertical vorticity downward against the flow; this was verified by Rotunno and Klemp (1985) and Walko (1993)]. If a Beltrami model is crudely assumed to represent the flow in a supercell (Davies-Jones and Brooks 1993), then vortex lines are coincident with streamlines and parcels flowing into the updraft at very low levels do not have significant vertical vorticity until they have ascended a few kilometers. Otherwise, argued Davies-Jones, abrupt upward turning of streamlines, strong pressure gradients, and large vertical velocities would be required next to the ground.

Davies-Jones (1982a,b) neglected baroclinic vorticity and suggested that the downdraft had the following roles in near-ground mesocyclogenesis: 1) tilting of horizontal vorticity by a downdraft produces vertical vorticity, 2) subsidence transports air containing vertical vorticity closer to the surface, 3) this air flows out from the downdraft and enters the updraft where it is stretched vertically, and 4) convergence beneath the updraft is enhanced by the outflow. Davies-Jones also showed kinematically that the flow responsible for tilting and concentrating vortex lines also tilts and packs isentropic surfaces, thus explaining observations of strong entropy gradients across mesocyclones near the ground.

c. Potentially relevant simulation results

Davies-Jones and Brooks (1993) showed that the vertical vorticity of air parcels descending in an RFD can be reversed during descent, from anticyclonic initially, to less anticyclonic, then to cyclonic in the lowest 50–125 m of their descent. As air subsides in the downdraft, vortex lines turned downward due to the barotropic “frozen fluid lines” effect (Helmholtz’s theorem), but with less inclination than the trajectories because horizontal southward vorticity was being generated continuously by baroclinity within the hook echo. Because of the geometry, vortex lines crossed the streamlines from lower to higher ones (with respect to the ground), and the barotropic effect served to turn the vortex lines upward even during descent. The baroclinic effect acted to increase horizontal vorticity further but did not control the sign of the vertical vorticity; thus, air with cyclonic vertical vorticity appeared close to the ground. As this air passed from the downdraft into the updraft, its cyclonic spin was amplified substantially by vertical stretching (Fig. 17).

Brooks et al. (1993, 1994) found that the formation of persistent near-ground rotation was sensitive to the strength of the storm-relative midlevel winds. When storm-relative midlevel flow was weak, RFD outflow undercut the updrafts and associated mesocyclones. When storm-relative midlevel flow was too strong, the cold pool was not oriented suitably for vorticity generation within the baroclinic zone immediately behind the updraft, which was found to be needed for the development of near-ground rotation in their simulations.

Wicker and Wilhelmson (1995) used a two-way interactive grid to study tornadogenesis. During a 40-min period, two tornadoes grew and decayed within the mesocyclone. Wicker and Wilhelmson’s Fig. 9 depicted a spiraling, asymmetric RFD associated with tornadogenesis. Their figure also indicated anticyclonic vertical vorticity on the opposite side of the RFD as the cyclonic vertical vorticity. Furthermore, the RFD contained low \( \theta_e \) values (\( \theta_e' \) was as small as \(-15\) K; \( \theta_e' \) was approximately \(-5\) to \(-8\) K in the hook echo).

Wicker and Wilhelmson found that parcels entered the mesocyclone from the RFD (Fig. 18) and descended...
from \(-500\) m, as Davies-Jones and Brooks (1993) had found. Furthermore, these parcels initially contained negative vertical vorticity; however, vertical vorticity increased to only weakly negative values, not to large positive values as in the simulations of Davies-Jones and Brooks. Trajectories entering the vortex have come from the hook echo and RFD region. [Adapted from Wicker and Wilhelmson (1995).]

Though numerical models have been instrumental in advancing our understanding of supercell dynamics, models may have limited utility in exploring the thermodynamic characteristics of RFDs, and the potential sensitivity of low-level vorticity intensification to these characteristics, owing to the unavoidable parameterization of microphysical processes. For example, three-dimensional simulations have not been able to produce the warm and moist RFDs that often have been observed near some strong tornadoes—the RFDs of simulated supercells almost invariably have large potential temperature and equivalent potential temperature deficits, as discussed in section 3c. However, some idealized simulations have suggested that tornadogenesis may be favorable if downdrafts are not too cold. Eskridge and Das (1976) proposed that a warm, unsaturated downdraft could be important for tornadogenesis; however, they did not specify whether the downdraft also could be cold, nor what the advantages of a warm downdraft over a cold downdraft were. Davies-Jones (2000b) recently has shown that an annular rain curtain can transport sufficient angular momentum from aloft to the ground to result in tornadogenesis in an idealized axisymmetric numerical model.\(^{16}\) No evaporation was permitted in the model; thus, no cold downdraft air was present. Furthermore, Leslie and Smith (1978) found that some vortices could not establish contact with the ground when low-level stable air was present, even if very shallow. Remarkably, Ludlam (1963) many years earlier had argued that “at least a proportion of the air that ascends in the tornado must be derived from the cold outflow; if this contains the potentially cold air from middle levels its ascent might be expected soon to impede if not destroy the tornado . . . it may be particularly important for the intensification and persistence of a tornado that some of the downdraft air may be derived from potentially warm air which enters the left flank of the storm at low-levels.”

Significant advances in our understanding of supercells no doubt have been made by numerical models. It is probable that some conclusions drawn from simulation results never could have been made from observations or theory alone. However, the author shares the view expressed by Doswell (1985): “The RFD’s role remains confusing with respect to tornadogenesis. Truly confirming evidence about the various aspects of numerical simulations awaits better observations, despite the compelling similarities between simulations and real storms.”

6. Recent observations from VORTEX

New radar and surface observations having unprecedented spatial and temporal resolution have been acquired by VORTEX and smaller, post-VORTEX field experiments. The findings of these recent operations that are pertinent to this review are highlighted below, with the exception of analyses of new surface data within hook echoes and RFDs, which will be presented by Markowski et al. (2002).

Recent radar observations, which include data obtained from both airborne and ground-based mobile Doppler radars (Jorgensen et al. 1983; Ray et al. 1985; Bluestein and Unruh 1989; Bluestein et al. 1995; Daugherty et al. 1996; Wurman et al. 1997), have resolved structures within hook echoes that were barely resolvable or unresolvable in the early radar studies of supercell storms. For example, in high-resolution (<100 m spatially) data of tornadoes presented by Wurman et

\(^{16}\) This simulation had some similarities with those conducted by Das (1983, unpublished manuscript), in which a precipitation-driven downdraft was imposed upon a wind field in which angular momentum increased with height. The downdraft was found to be able to transport sufficient vorticity from aloft to result in tornadogenesis.
al. (1996), Wurman and Gill (2000), and Bluestein and Pazmany (2000), the echo-free holes that were only marginally resolved by Garrett and Rockney (1962) were well resolved (Figs. 19 and 20). The images presented by Bluestein and Pazmany (2000) even begin to marginally resolve structures, possibly subvortices, within the echo-free hole itself. Moreover, the radar reflectivity depictions “looked like a tropical cyclone, with concentric inner bands and outer spiral bands” (Bluestein and Pazmany 2000). Hook echoes as narrow as 100 m or less have been detected (Wurman et al. 1996; Bluestein and Pazmany 2000), perhaps implying that precipitation loading and evaporative cooling within the hook echoes of some storms may not be the most significant effects in driving the associated RFDs, at least at low levels.

Just as the early multiple-Doppler radar analyses of the 1970s and 1980s revealed, radar observations obtained in the last decade also have detected vorticity doublets straddling the hook echoes of tornadic storms (Rasmussen and Straka 1996; Wurman et al. 1996; Straka et al. 1996; Bluestein et al. 1997; Dowell and BLuestein 1997; Dowell et al. 1997; Wakimoto and Liu 1998; Wakimoto et al. 1998; Wurman and Gill 2000; Ziegler et al. 2001) and nontornadic storms (Gaddy and Bluestein 1998; Blanchard and Straka 1998; Wakimoto and Cai 2000; Bluestein and Gaddy 2001). These vorticity doublets could be evidence that RFDs are involved in a downward displacement of initially quasi-horizontal vortex lines, perhaps necessarily transporting rotation to the surface during tornadogenesis, as many previous investigators (e.g., Ludlam, Fujita) have conjectured.

Using surface observations obtained from automobile-borne sensors (Straka et al. 1996), Rasmussen and Straka (1996) documented a relatively warm RFD south of the Dimmitt, Texas, tornado. In the same case, which was during VORTEX, the hook echo was collocated with the surface divergence maximum, implying an association between the hook echo and (at least) a low-level downdraft, as also had been suggested by numerous predecessors.

In two other VORTEX storms, Wakimoto et al. (1998) and Wakimoto and Cai (2000) concluded that the occlusion downdraft was driven largely by the reversal of the vertical gradient of dynamic pressure, owing to increasing vorticity at low levels. In the supercell documented by Wakimoto et al. (1998; the 16 May 1995 VORTEX storm), it was found that the precipitation-loading forcing of the occlusion downdraft was an order of magnitude less than the forcing provided by the nonhydrostatic vertical pressure gradient. [In a different case, Carbone (1983) previously had suggested that precipitation loading may contribute to occlusion downdraft genesis.] In the 12 May 1995 VORTEX storm studied by Wakimoto and Cai (2000), a thermodynamic retrieval indicated that the occlusion downdraft was associated with a warm core.

Perhaps the most remarkable observational finding during the last 10 years is that the differences between tornadic and nontornadic supercells may be subtle, if
Fig. 20. Radar reflectivity fields (dBZ) associated with a hook echo on 15 May 1999 (left) early in the life of a tornado, (center) during the mature stage, and (right) during the dissipation stage. [From Bluestein and Pazmany (2000).]

even distinguishable, even in dual-Doppler radar analyses of the wind fields just prior to tornadogenesis. Blanchard and Straka (1998) documented a mobile radar signature of a spiraling hook echo in a nontornadic supercell having an appearance similar to those that have been associated with tornadic supercells (e.g., similar to the radar image in Fig. 20). Perhaps such near-ground circulations are more common in nontornadic supercells than previously believed. Trapp (1999) and Wakimoto and Cai (2000) also documented circulations in nontornadic supercells at levels close to the ground. Trapp found that the low-level mesocyclones associated with tornadogenesis had smaller core radii and were associated with more substantial vorticity stretching than those associated with tornadogenesis “failure.” Based on a comparison of pseudo-dual-Doppler analyses, Wakimoto and Cai concluded that the “only difference between the Garden City storm and Hays storm (the 16 May 1995 and 12 May 1995 VORTEX storms) was the more extensive precipitation echoes behind the rear-flank gust front for the Hays storm.”

7. Concluding remarks

The association between hook echoes, RFDs, and tornadogenesis has been well documented for nearly 50 years; however, the precise dynamical relationship still is not known today. The analysis of the three-dimensional wind structure of supercells afforded by Doppler radar, along with speedy increases in the feasibility of numerical cloud modeling, led to relatively rapid gains in knowledge of the recurrent storm structures and evolution associated with supercells. Within 30 years of the first radar image of a hook echo, we knew that the most damaging tornadoes were associated with supercells, we knew about the existence of the parent circulations of tornadoes (mesocyclones), we developed an understanding of the dynamics of midlevel storm rotation and storm propagation, radar and visual features common to supercells were well documented, and downdrafts were recognized as being important in tornadogenesis. Yet in the decades that followed the period of rapid advances, no breakthroughs emerged with respect to the role of the RFD in tornadogenesis. In fact, it is debatable whether we can better anticipate tornadogenesis within supercell storms today than we could 20 years ago.

If the hook echo and its associated RFD truly are critical to tornadogenesis, as hypothesized for many years, then perhaps significant gains in understanding will not be possible until more spatially and temporally detailed observations of this region can be made, in addition to numerical simulations with more realistic
representations of entrainment and microphysical processes. It is believed that some of the important outstanding questions include

- What are the dominant forcings for RFDs, as a function of location within the RFD and stage in storm evolution?
- How do the dominant RFD forcings vary across the spectrum of supercell types (e.g., nontornadic vs tornadic, low-precipitation vs heavy-precipitation storms)?
- How do the thermodynamic and microphysical characteristics of hook echoes and RFDs vary across the supercell spectrum, and why?
- How does the large-scale environment affect RFD characteristics?
- Is the tornadogenesis process sensitive to the thermodynamic and microphysical properties of RFDs?
- What is the role of the RFD in tornadogenesis, and does the hook echo have an active role?

A number of direct observations have been reviewed herein; however, these observations have been relatively scarce and often have been simply fortuitous. A new mobile surface observing system (Straka et al. 1996), introduced in 1994 for VORTEX, recently has collected the largest number of in situ measurements within supercells to date. Analyses of these spatially and temporally dense “mobile mesonet” observations within hook echoes and RFDs will be presented in a companion paper (Markowski et al. 2002), and these data may begin to shed some light on at least a couple of the questions posed above.

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REFERENCES


Hane, C. E., and P. S. Ray, 1985: Pressure and buoyancy fields derived


