Kinematic, Thermodynamic, and Visual Structure of Low-Reflectivity Microbursts

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ABSTRACT

On 9 July 1987, a series of low-reflectivity microbursts were studied over Colorado using dual-Doppler analyses, cloud photogrammetry, and in situ measurements collected by aircraft. These types of wind-shear events are particularly hazardous to the aviation community since the parent cloud and pendant virga shafts appear innocuous. The microburst downdrafts are shown to develop at the location where the virga shafts are, visually, the lowest and opaque. As the downdraft intensifies, sublimation and evaporation (to a smaller extent) rapidly deplete the hydrometeors and result in a shift of the axis of maximum negative vertical velocities into a relatively low reflectivity and transparent region of the virga shafts. Comparisons with weak downdrafts or null cases reveal that the maximum radar reflectivities within the parent clouds for the two cases are comparable; however, the microburst storm consistently exhibits a larger horizontal area encompassed by the 10-dBZ contour at midlevels prior to downdraft formation.

1. Introduction

Although the downdraft has long been recognized as one of the fundamental flows associated with precipitating convective clouds, there are many facets that are not fully understood (Knupp and Cotton 1985). An example is a class of very strong downdrafts and accompanying outflows, called microbursts. These wind-shear events have been the focus of a considerable amount of research and have been identified as a causal factor in a number of aircraft accidents (Fujita 1976; 1986; Fujita and Byers 1977; Fujita and Caracena 1977). The small spatial and temporal scales of this phenomenon make it extremely difficult to study and, most importantly, forecast in real time.

It has been shown that microbursts can develop under certain favorable environmental conditions. Brown et al. (1982) and Wakimoto (1985) have illustrated that strong outflows will form when the environment exhibits a deep (≈3 km), dry-adiabatic subcloud layer, consistent with the earlier speculations by Braham (1952) and Krum (1954). Numerical calculations by several investigators (Hookings 1965; Kamburova and Ludlam 1966; Srivastava 1985, 1987; Proctor 1989) suggest that the maintenance of a downdraft by the evaporation of falling precipitation is a function of drop size, rain intensity, and downdraft speed. One of their conclusions is that if the environmental lapse rate was approximately equal to the dry-adiabatic lapse rate, then the rates of evaporation place little restriction on downdraft magnitude, and even in light precipitation, strong downdrafts may be generated. These results have been confirmed with radar and aircraft observations of virga shafts from weakly precipitating cloud systems (McCarthy and Serafin 1984; Wilson et al. 1984; Fujita 1985; Wakimoto 1985; Kessinger et al. 1986; Mahoney and Rodi 1987). Striking in the results presented by Wilson et al. (1984) and Hjelmfelt (1988), based on data collected during the Joint Airport Weather Studies (JAWS, McCarthy et al. 1982), is that the strongest microburst during the project (approximately 50 m s⁻¹ velocity differential) was associated with only a 25-dBZ echo at a height of 500 m AGL.

These types of microbursts are often referred to as "low-reflectivity" (Kessinger et al. 1986; Hjelmfelt et al. 1989; Roberts and Wilson 1989) or "dry" (Brown et al. 1982; Fujita 1985; Wakimoto 1985; Mielke and Carle 1987; Proctor 1989) microbursts and are hazardous to aircraft since the parent cloud and pendant virga shafts appear innocuous. Based on data collected during the JAWS project, Kessinger et al. (1986) have shown that the average velocity differential for low-reflectivity microbursts observed with Doppler radar

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is 24 m s\(^{-1}\). The clouds are typically shallow (3–6 km deep) with a high cloud base (3–5 km AGL). Forecasting techniques for these types of microbursts based on single-Doppler radar (Roberts and Wilson 1989) and thermodynamic soundings (Wakimoto 1985) have been established to help warn the general public and the aviation community.

A study by Mahoney and Rodi (1987) was one of the first documenting the kinematic, thermodynamic, and microphysical characteristics of low-reflectivity microbursts based almost entirely on aircraft data. Their main conclusion was that subcloud evaporation was the dominant mechanism for downdraft generation. Detailed multi-Doppler data on low-reflectivity microbursts was presented by Hjelmfelt et al. (1989); however, owing to the concentration of radar scanning at low levels, the complete storm structure was not presented. Combined with two-dimensional modeling studies, they suggest that precipitation loading of graupel/hail and subcloud melting and evaporation were important. Other general kinematic features of the low-level structure of these types of microbursts have been discussed by Wilson et al. (1984) and Hjelmfelt (1988).

On 9 July 1987 a comprehensive dataset was collected on a series of dry microbursts during the Convective Initiation and Downburst Experiment (CINDE, Wilson et al. 1988) near Denver, Colorado. This case is unique for several reasons

1) Multi-Doppler radar data were collected over the entire depth of the storm and over the life cycle of several microbursts. In addition, several cases of weak downdrafts (referred to as null cases) were also analyzed and compared with the microburst events. This type of comparison has not been attempted in the past for low-reflectivity storms. Potts (1989) examined null events from moderate- and high-reflectivity storms.

2) A series of photographs were taken of the parent cloud documenting the visual characteristics of the virga shafts associated with the microbursts. These pictures combined with other observational datasets provide for a more quantitative interpretation of these pendant precipitation shafts.

3) High- and low-level aircraft penetrations recorded the thermodynamic and microphysical characteristics of the downdraft.

Perhaps the most important applied result from this case is item 2. It is critical to determine if there are identifiable features associated with virga shafts that can be used as a warning to pilots of impending strong wind-shear events. Section 2 discusses the data and analysis methods. The environmental conditions are presented in section 3. The aircraft and radar analyses are shown in sections 4 and 5, respectively. Section 6 discusses the forcing mechanisms based on numerical simulations using a one-dimensional microphysical model. Section 7 presents analyses of storms that did not produce microbursts on this day (null cases), and a summary and discussion of the results is presented in section 8.

2. Data and method

During the summer of 1987, the multiagency field program CINDE was conducted near Denver, Colorado, by the National Center for Atmospheric Research (NCAR) and the National Oceanic and Atmospheric Administration (NOAA) with participation from several universities. Part of the measurement network included two X-band radars (NOAA-C and NOAA-D) of the NOAA Wave Propagation Laboratory (WPL), the NCAR C-band CP-3 radar, 46 NCAR Portable Automated Mesonet (PAM-II) surface stations, the permanent array of 22 surface stations maintained by the NOAA Forecast Systems Laboratory (FSL), the NCAR and University of Wyoming King Air research aircraft, University of North Dakota Citation research aircraft, 5 NCAR Cross-chain Loran Atmospheric Sounding Systems (CLASS), 2 NCAR and 1 NOAA mobile sounding systems, and photographic equipment. Locations of the ground-based facilities are shown in Fig. 1. The reader is referred to Wilson et al. (1988) for more information.

On 9 July 1987, a virga line that subsequently produced a series of low-reflectivity microbursts moved into an excellent location for dual-Doppler observations (see Fig. 1). Table 1 lists parameters for each of the Doppler radars. The NOAA-C and NOAA-D radars were collecting data using narrow sector (35°–50°), plan position indicator (PPI) volume scans. Volumes were collected at 130–140-s intervals on two nonmicroburst-producing storms from 1621 to 1630 MDT (hereafter all times in MDT) and at 120–150-s intervals on a microburst-producing storm from 1645 to 1705. The NOAA radars were limited by their data collection system to processing less than 100 range gates over a small PPI sector. For this reason, an entire storm could not be sampled during a volume scan and scanning had to concentrate on a specified area of interest. Therefore, data from the NCAR CP-3 Doppler radar were used to generate histories of the storms of interest since it continuously operated in 360° surveillance mode.

Radar data were edited using the NCAR Interactive Doppler Editing System (Oye and Carbone 1981). Contamination from ground clutter, low signal-to-noise threshold, and “second trip” echoes was removed. Aliased radial velocity values were corrected. Once edited, the data were interpolated to Cartesian space using a Crossman (1959) distance-dependent weighting scheme where the \( X, Y, \) and \( Z \) radii of influence were 0.25 km. A storm-motion vector of 15.2 m s\(^{-1}\) toward 55° was applied during the interpolation process. One grid domain was used for all dual-Doppler analyses even though each analysis filled only part of it. The analysis domain had \( X, Y, Z \) dimensions of 20 km.
$20 \text{ km} \times 7.5 \text{ km}$ with 0.25-km grid spacing for all dimensions. The first analysis level was 0.06 km AGL. The minimum resolvable wavelength is approximately 1.5 km given the characteristic dimensions of the radar data (Carbone et al. 1985).

After the radar data were interpolated to Cartesian space, the NCAR CEDRIC radar analysis software package (Mohr et al. 1986) was used to synthesize the three-dimensional wind components using the two radar equations (Miller and Anderson 1991). Initial estimates of $u$, $v$ were smoothed with a two-pass Shuman (Shuman 1957) filter before calculation of the horizontal convergence field. Downward integration of the horizontal convergence by means of the anelastic continuity equation was performed to calculate the initial vertical velocity estimates. The hydrometeor fall speeds were estimated from radar reflectivity (Joss and Waldvogel 1970) and were corrected for the effects of air density (Foote and duToit 1969). Downward integration proceeded from a boundary condition of 0.0 m s$^{-1}$ applied 1 km above the storm top. Convergence to a final solution was achieved through iteration. The resultant $u$, $v$, $w$ wind components were variationally adjusted with lower and upper boundary conditions of 0.0 m s$^{-1}$ applied at the ground and 1 km above echo top. Since additional winds could be determined in the boundary layer using upward integration, the process was repeated and the horizontal wind components were used to “fill” areas in the lowest analysis regions for display purposes only. The vertical component was not “filled.”
Following statistical techniques described by Doviak et al. (1976) and Kessinger et al. (1987), uncertainty in the wind fields is estimated based on the two-radar geometry. Random error values of 1 m s\(^{-1}\) are assumed for the radial velocity estimates from each radar and the terminal fall-speed estimate. Downward integration from a boundary condition of 0.0 m s\(^{-1}\) for vertical velocity is used. Error estimates for the horizontal divergence are approximately \(1 \times 10^{-2}\) to \(3 \times 10^{-3}\) m s\(^{-1}\) over the analysis domain. For the vertical velocity, error estimates may vary from 1 to 5 m s\(^{-1}\).

Figure 1 also schematically shows the geometry for a series of photographs taken from the NOAA-C radar site of the parent cloud, virga shafts, and the blowing dust caused by the microbursts. Photogrammetric techniques require knowledge of the camera location and the azimuth angles to landmarks visible in the photographs. In the analyses presented in section 5, rows of trees will be evident on the horizon that provided the landmarks for this purpose. Once these angles are known, an entire azimuth and elevation angle grid can be constructed and superimposed on the photograph. A general discussion of photogrammetry techniques is given by Holle (1986).

3. Environmental conditions

The synoptic forcing on 9 July 1987 was relatively weak. At upper levels, a trough was located over the Pacific Northwest resulting in southwesterly flow over Colorado. The thermodynamic soundings launched from Denver at 0600 and Hudson at 1705 are shown in Fig. 2. (See Fig. 1 for site locations.) The Hudson sounding is believed to be representative of the ambient conditions since it was, geographically, the closest to the microburst outflows and was launched within 20 min of the first dual-Doppler analysis of the microburst-producing storm. The vertical wind profile shows a slight backing of the winds during the day as the trough slowly approaches from the west.

![Fig. 2. The Denver (DVR) and Hudson (HUD) soundings launched at 0600 and 1705 MDT, respectively. Both launch sites are shown in Fig. 1.](image)

The temporal evolution of the soundings in Fig. 2 is consistent with the schematic model proposed by Brown et al. (1982) and Wakimoto (1985) shown to be conducive for low-reflectivity microbursts. Prominent is the deep, dry-adiabatic subcloud layer (3–4 km) and dewpoint depressions as large as 30°C near the surface and decreasing to near 0°C at cloud base. Based on a statistical sample of 186 wind-shear events, Wakimoto (1985) showed that microburst activity normally begins in the afternoon with primary and secondary peaks at 1500 and 1800, respectively. On 9 July the first of numerous reports of microbursts within the CINDE network was at approximately 1515 MDT.

In an attempt to understand the temporal evolution of the boundary layer on this day, a time series of the soundings launched from Denver is shown in Fig. 3.

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<th>Table 1. Doppler radar parameters.</th>
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<td><strong>NOAA C and D</strong></td>
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<td>Site longitude</td>
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<td>Feedhorn elevation (km MSL)</td>
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| **NCAR CP-3**                     |
| Wavelength (cm)                   | 5.45 |
| Beamwidth                         | 1.11°|
| Pulse duration (µs)               | 1.0  |
| PRF (Hz)                          | 1250 |
| Average transmitted power (dBm)   | 58.5 |
| System gain (dB)                  | 43   |
| Approximate system noise power (dBm)| −109|
| Gate spacing (m)                  | 150  |
| Number of gates                   | 512  |
| Elevation angles used in scanning strategy | 0.4°, 0.9°, 1.7°, 2.6°, 3.5°, 4.5°, 5.5°, 6.5°, 7.5°, 8.5°, 10.0°, 12.0°, 15.0° |
| Site latitude                     | 39°45′50″N |
| Site longitude                    | 104°43′58″W |
| Feedhorn elevation (km MSL)       | 1.67 |

**Table 1. Doppler radar parameters.**
4. The microburst-producing storm and aircraft analyses

The general structure of the storm systems on 9 July are consistent with the comprehensive results presented by Kessler et al. (1986) and Roberts and Wilson (1989). Tracks of all storms passing through the northern half of the CINDE network from 1500 to 1840 MDT are shown in Fig. 4 and are based on CP-3 reflectivity data at 5.5° elevation. (Storms were identified by exhibiting reflectivities greater than or equal to 10 dBZ at this elevation angle.) Storms that produced microburst outflows are shown with solid lines; storms that did not are shown with half-toned lines. This figure shows that many storms were present on this day; microburst occurrence was widespread. The storms that are featured in this paper are numbered 21, 22, and 23a. Using the Wilson et al. (1984) criteria for defining microburst outflows (Doppler radial velocity divergence from peak to peak greater than or equal to 10 m s⁻¹ over 4 km), storms 21 and 22 did not produce outflows of sufficient strength to be classified as microbursts; storm 23a produced several intense microbursts.

A PPI display of storm 23a, the primary storm of interest, is shown in Fig. 5. Striking in this figure are the relatively low reflectivity values accompanying the storm, which was moving at 15.2 m s⁻¹ toward 55°. The microbursts generated by the storm were intense, with a maximum differential velocity at the surface measured by the NOAA-C radar in excess of 30 m s⁻¹. The surface data recorded by PAM stations 8 and 9 in Fig. 5 imply a pattern of surface divergence associated with the microburst outflow, and a time series from the former station is shown in Fig. 6. The familiar peak in surface wind speed at 1654 MDT is associated with the microburst outflow (e.g., Fujita 1985). Although there is a slight warming in the potential temperature θ when the outflow reaches the station, there is a substantial decrease in equivalent potential temperature θe. These two observations suggest that the microburst was warm and very dry.

Two low-level penetrations by the Citation are shown in Fig. 5 at 1651:12 and 1653:21 MDT, with the details of the former shown in Fig. 7. Figure 7 illustrates that the Citation penetrated the microburst downdraft and horizontal rotor circulations (Fujita 1985; Kessler et al. 1988; Parsons and Kropfli 1990). Note that the increased head winds and updrafts associated with the rotor caused the Citation to rise to over 400 m AGL at 1650:21 followed by a loss of approximately 200 m in altitude while penetrating the downdraft. This scenario is similar to several commercial aircraft accidents (Fujita 1985). The Citation

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**Fig. 3.** Time series of soundings launched from Denver (DVR). The location of cloud base and the time when the microburst activity was documented within the CINDE network are shown. Hourly surface reports of temperature and pressure from the Denver National Weather Service office were used to supplement the analyses.

**Fig. 4.** Tracks for all storms in the northern half of the CINDE network from 1500 to 1840 MDT based on CP-3 reflectivity data greater than or equal to 10 dBZ at 5.5° elevation. Storms that produced microburst outflows are shown with solid black lines; storms that did not are shown with half-tone lines. Storm motion was toward the northeast.
data show little change in $\theta$ and $\theta_E$ inside and outside of the microburst downdraft, which may be attributed to mixing within the horizontal rotor circulation. The second leg by the Citation shown at 1653:21 in Fig. 5 did not penetrate a microburst downdraft; however, the horizontal winds do reveal the overall divergence away from the storm.

Also shown in Fig. 5 is the location of an earlier higher-level penetration ($\approx$ 3 km AGL) through the same microburst downdraft by the Wyoming King Air. The track shown is one of several penetrations of the downdraft associated with storm 23a (Rodi and Fankhauser 1989). This track has been time–space adjusted to agree with the radar echo pattern recorded approximately 1 min after the penetration. At this height, the convergence of the south-southwest and southwest flow observed on this south to north flight track across the echo is evident. This midlevel convergence has been noted as a predictor for microburst development by Roberts and Wilson (1989). The plot of the vertical motion (Fig. 8) shows that the maximum microburst downdraft speed was $-12.8$ m s$^{-1}$ with the number of precipitation particles observed by a Particle Measuring Systems (PMS) 2D-P probe in excess of 5 L$^{-1}$. This probe measures particle sizes larger than 0.2 mm (Mahoney and Rodi 1987). Particle concentrations were also recorded by the Citation aircraft; however, at these lower levels no precipitation particles were detected. Mahoney and Rodi (1987) did not detect any hydrometeors during a 1.7-km AGL penetration of a low-reflectivity microburst.

The relatively cold downdraft air in Fig. 8 is revealed by the potential temperature and, to some extent, the equivalent potential temperature. The change in the thermal structure in Figs. 6 and 8 from a cold downdraft aloft to one that is slightly warmer than the environment at low levels has been explained by Srivas-
starting at 1645:56 and ending at 1656:48 MDT are shown in Fig. 10. The surface analyses show the individual microburst outflows labeled sequentially with letters from A to D. Storm 23a produced a series of microbursts along a line, similar to cases documented by Hjelmfelt (1987) and Hjelmfelt et al. (1989). Each new microburst reached the surface toward the northeast of the previous one and eventually produced an elongated area of surface divergence whose major axis was along the storm motion path. The strongest surface divergence exceeded $20 \times 10^{-3}$ s$^{-1}$ for several of the microbursts shown in Fig. 10.

Cross sections are shown at 3.56 km AGL since this was identified as the level of maximum convergence into the downdraft, which at times exceeded $-10 \times 10^{-3}$ s$^{-1}$. Roberts and Wilson (1989) have discussed the usefulness of detecting this feature with a single-Doppler radar to nowcast microburst events. The convergence almost extends the entire length of the line and is nearly two-dimensional, consistent with the findings from Hjelmfelt et al. (1989). Noteworthy in Fig. 10 is the suggestion that the convergence is occurring at the northern edge of the echo along the maximum gradient of radar reflectivity and does not seem to be collocated with a maximum reflectivity core [Roberts and Wilson (1989) also noted this feature

tava (1985). He showed that the thermal buoyancy could be negative even though the downdraft is warmer than its environment since the descending air was much drier and hence virtually cooler than the environment.

Fortunately, the King Air was equipped to record two-dimensional imaging probe data (Fig. 9). This is believed to be the first time that the microphysical characteristics of low-reflectivity microbursts have been clearly defined. This is important in light of the recent numerical work examining the sensitivity of strong downdrafts to different precipitation types (Krueger et al. 1986; Proctor 1989). Based on the results from Heymsfield and Musil (1982), the images shown in Fig. 9 are highly suggestive of rimed ice crystals and rimed aggregates (see their Figs. 16 and 17). This information is used in the numerical simulations discussed in section 6.

5. Radar analysis

a. Dual-Doppler synthesis

Horizontal cross sections of the storm-relative flow at 0.06, 2.31, and 3.56 km AGL for five scan volumes

FIG. 7. Time series of the low-level penetration of the Citation shown at 1651:12 MDT in Fig. 5. Vertical and horizontal wind speed, wind direction, potential and equivalent potential temperature, and height of the aircraft are plotted. Arrows and streamlines in the middle of the figure represent the total vertical motion and the component of horizontal wind along the flight track. Approximate horizontal length scale is drawn on the figure.

FIG. 8. Time series of the high-level penetration of the Wyoming King Air at 1646:48 MDT shown in Fig. 5. Vertical and horizontal wind speed, wind direction, particle concentration, potential and equivalent potential temperature, and height of the aircraft are plotted. Approximate horizontal length scale is drawn on the figure.
associated with high-reflectivity storms.] This observation must be viewed with caution since it develops after the microburst reaches the surface. When the downdraft first initiates at the northeast end of the elongated echo there does seem to be a stronger tendency for the convergence to be near the maximum reflectivity values (note the 1645 and 1652 MDT analysis times). The observation of convergence along the reflectivity gradient is also apparent in Hjelmfelt et al. (1989) (see their Fig. 5), but it is never discussed.

The 2.31-km level was also chosen in Fig. 10 since this was the height that a pronounced radar reflectivity “bright band” (Battan 1973) was noted. There is relatively weak convergence at this level, nearly centered along the strongest reflectivities. Vertical cross sections along lines $AB$ and $CD$ shown in Fig. 10 are presented in Fig. 11. The cross section of microburst B was chosen since this was the downdraft that the Wyoming King Air and Citation penetrated. (The King Air flew near the level of the maximum convergence, just below the downdraft origin in Fig. 11.) The downdraft development on the northern edge of the echo is prominent. Given the fact that the southwesterly environmental winds shown in Fig. 2 are nearly parallel to the elongation of the echo, this preference for the north side is unclear.

The pronounced vertical depression of the echo at the first analysis time in Fig. 11 is probably the result of a rapid depletion (hypothesized to be sublimation) of hydrometeors as the downdraft descends. It should be noted that this mechanism for the formation of the echo depression appears to be contradicted by existence of weak updrafts between 4.5 and 6 km AGL. However, these small vertical motions (primarily less than 1 m s$^{-1}$) are within the error range discussed in section 2. It is believed that the downdrafts probably extend to the echo top.

The scenario discussed in Fig. 11 is supported in Fig. 12 by a series of vertical cross sections following the downdraft associated with microburst C. The evolution is similar to that shown in Fig. 11; however, the sequence of events starting from an initial high-reflectivity core into a weak-echo trench and a pronounced bright band is clearly illustrated. These cross sections are reminiscent of several shown by Fujita (1985) (see his Figs. 6.12, 6.35, and 6.38) using single-Doppler radar information. One other reflectivity hole associated with a virga line storm that produced microbursts was noted within the CINDE network by CP-3 surveillance scans beginning near 1800 on this day.

As the core of microburst B descends, the snow aggregates, shown in Fig. 9, melt and form a bright band.
This suggests that a combination of ice loading and sublimation alone drive the early microburst intensification. Details of this microphysical evolution of the downdraft are discussed in section 6. The horizontal divergence near 1 km accompanying the microburst outflow is evident at this time.

The microburst has impacted the surface by the second volume time (1647:59–1650:09) in Fig. 11. The reflectivities along the downdraft axis have weakened considerably at this time. One of the horizontal rotor circulations that was penetrated at 200–400 m AGL by the Citation aircraft (Fig. 7) is clearly depicted.

Figure 13 is a plot of the $\theta_E$ based on the Hudson sounding launched at 1705 MDT. The downdraft can be seen to originate just above the freezing level and below the level of minimum $\theta_E$. Recall from Fig. 6 that the values of $\theta_E$ recorded by station 8 during the passage of the outflow were approximately 336 K. It is believed that these temperatures originate from very near the surface. Unfortunately, the moisture profile from the sounding at low levels was not reliable and has been eliminated in Fig. 13. However, the sounding shows that the 335.5-K isopleth is near 1 km, which is consistent with a low-level origin of the downdraft air reaching the surface. Numerical simulations by Krueger (1988) also support this. The vertical displacement between the 0°C isotherm and the bright band that is evident in Fig. 13 will be discussed in more detail in sections 5b and 6.

b. Cloud photogrammetric analyses

As previously mentioned, a series of photographs were taken from the NOAA-C site of storm 23a, pendant virga shafts, and the blowing dust at the surface. An elevation and azimuth angle grid was superimposed on the pictures using photogrammetric techniques. Once this grid was determined, it was possible to superimpose vertical cross sections through the center of the microburst downdraft of radar reflectivity and storm-relative Doppler winds (Fig. 14). The justification for performing this type of analysis is to determine the structural relationship of the visual virga shafts with the microburst downdraft. These results may prove invaluable by providing a visual clue that can be used by airline pilots to avoid wind-shear events.

The first analysis time is presented in Fig. 14a. The location of cloud base and the environmental melting level as determined from the Hudson sounding are shown on this and subsequent figures. The location of the King Air penetration through the virga shafts associated with microburst B is also marked. A relatively transparent region devoid of hydrometeors is labeled as a virga hole on the figure.

The radar reflectivity cross section indicates relatively low values at the location of the virga hole. Also prominent is a well-defined bright band located at the visual termination of the virga shafts. This appears to be the first observational evidence illustrating that the end of a virga shaft marks the location where hydrometeors melt. This observation is consistent with optical thickness calculations by Frazier and Bohren (1992). They hypothesize that the abrupt change of the visual characteristics at the end of a virga shaft could not be caused by evaporating rain as commonly assumed by meteorologists. A more plausible precipitation type was melting snowflakes.
In the present case the brightband level is approximately 1 km below the environmental 0°C level. Battan (1973) states that in most cases the center of the bright band is generally from about 100 to 400 m below the 0°C isotherm. Results from experiments by Matsuo and Sasyo (1981a,b) and Rasmussen and Pruppacher (1982) have shown that melting is delayed when an ice particle falls in subsaturated air owing to cooling caused by sublimation. Therefore, melting may begin at environmental temperatures considerably warmer than 0°C.

Although not visually apparent, the radar reflectivities suggest that a few raindrops nearly reach the surface within the downdraft. This is not contradictory to the Citation observations that no hydrometeors were detected by the 2D-P probe at 200 m AGL.

The radar beamwidth is approximately 0.8° and results in a pulse volume average over about 300 m in the vertical. Since the lowest scan angle was 0.2°, particles from as high as 220 m could contribute to the observed reflectivities.

The decaying stage of microburst A and mature stage of microburst B are shown in the dual-Doppler wind field in Fig. 14a. Microburst B develops at the location where the virga shafts are, visually, the lowest. The latter observation is suggestive of a significant amount of hydrometeors, which is confirmed by the slightly higher reflectivities in this region. This is also the region where the bright band is the strongest and most clearly defined. There is agreement within 1−2 m s\(^{-1}\) between the wind measurement from the King Air and the dual-Doppler winds in the same area.
Microburst B has reached the surface in the next scan volume (Fig. 14b). The outflow is visually apparent as blowing dust in the shape of a curl, as documented by Fujita (1985) (see his Fig. 2.7). An enlargement of the lower end of the virga shafts and the blowing dust for this time is shown in Fig. 15. The dust reaches a height of 700 m. A weak-echo trench has formed along the downdraft axis at this time. A combination of sublimation and evaporation is hypothesized to be responsible for depleting the hydrometeors in this region. Figure 15 illustrates that the temporal evolution of the virga hole could make identification of the microburst downdraft difficult for airline pilots.

The Citation makes its penetration through the decaying stage of microburst B at 1651 (shown in Fig. 14c). There is good agreement between the Doppler and Citation wind data shown on the figure. A new microburst, labeled C, has developed and is situated in the visually opaque and highest reflectivity area of the virga shafts, similar to microburst B.

In Fig. 14d, microburst C reaches the surface and a reflectivity trench develops again along the downdraft axis. Slightly toward the east, microburst D begins to descend in the visually opaque and lowest area of the virga shafts. This process is repeated in Fig. 14e with the formation of microburst E.

The analyses in Fig. 14 strongly suggest that the microburst downdraft is visually associated with the opaque and lowest pendant virga shafts during the formative stages. However, the rapid depletion of hydrometeors owing to sublimation and evaporation that occurs as the microburst descends toward the surface may produce a
Fig. 14. Photographs of the microburst storm 23a superimposed on vertical cross sections through the center of the microburst downdrafts depicting radar reflectivity and storm-relative dual-Doppler winds from NOAA-C and NOAA-D. The locations and measured winds from the Wyoming King Air and Citation penetrations are shown in the top of panels (a), (c), and (d). (The lengths of the bold arrows representing these measured winds are consistent with the dual-Doppler winds.) The cloud base and the environmental melting level determined from the Hudson sounding are indicated in the figures. The location of the bright band is also indicated. The height grid is valid at the location of the microburst downdrafts. The dashed box in (b) is enlarged in Fig. 15.
relatively transparent and low-reflectivity area of the virga shafts (e.g., Fig. 14b). This latter time is when an aircraft at low altitudes during takeoff and landing is most susceptible to wind shear. A similar reduction in reflectivity values within a high-reflectivity microburst over Colorado was noted by Lee et al. (1992).

Vertical cross sections of vertical velocity, divergence, and reflectivity for microburst B are shown in
intensification. The horizontal convergence centered below 4 km AGL and the low-level divergence associated with the microburst outflow at the surface are apparent.

6. Analysis of forcing mechanisms with a one-dimensional model

The forcing mechanisms for low-reflectivity microbursts have been examined previously by several investigators. Roberts and Wilson (1989) only speculated, based mainly on single-Doppler radar data, what mechanisms could be forcing the microburst. Mahoney and Rodi (1987) had detailed aircraft penetrations through a microburst; however, even though they suspected that graupel was present, their calculations of cooling rates were only for liquid drops. Hjelmfelt et al. (1989) ran several two-dimensional cloud model simulations and concluded that loading by graupel/hail was important along with the subsequent melting and evaporation of rain. Results from the current study do not suggest that hail was present, and the analysis presented in Fig. 16 suggests that much of the acceleration of the microburst occurs well before melting takes place. Proctor (1989) appears to be the first to suggest, based on numerical simulations, that sublimation of snowflakes can be the dominant driving mechanism for low-reflectivity microbursts. This microburst case with its unusually detailed Doppler and aircraft measurements appears to be the ideal observational dataset to test Proctor’s hypothesis.

To better quantify the microphysical forcing mechanisms leading to downdraft intensification for the microburst storm, the one-dimensional time-dependent model of Srivastava (1985, 1987) was applied. The

Fig. 14. (Continued)
Fig. 15. Enlargement of the end of the virga shafts and the blowing dust from microburst B shown in Fig. 14b.

Fig. 16a. The rapid acceleration of the downdraft occurs between cloud base and the bright band. This strongly suggests that sublimation of the snow aggregates shown in Fig. 9 plays an important role in the microburst
model contains detailed microphysics for both water and ice substances and has equations for calculating raindrop evaporation, melting and sublimation of ice particles, raindrop and ice concentrations, mixing ratio of ice and water substance, thermodynamic energy, and vertical air velocity. At the top of the downdraft, the pressure, temperature, relative humidity, vertical air velocity, and ice size distribution are specified while the bottom of the downdraft is considered open. The environment is characterized by the vertical profiles of temperature and moisture. Calculations are made on the evolution of the downdraft in the subcloud layer.

By considering the maximum radar reflectivity measurements of 22 dBZ in conjunction with two-dimensional particle images in Fig. 9, a monodispersed size distribution was chosen for a snowflake diameter of approximately 1.72 mm (melted diameter of about 0.8 mm). The particle concentration based on the aircraft measurements was set at $3 \times 10^{-5}$ cm$^{-3}$. Since the images in Fig. 9 suggest the presence of rimed aggregates, the ice-particle density was fixed at 0.1 g cm$^{-3}$, which represents an average of the densities of unmixed aggregates [i.e., 0.02-0.1 g cm$^{-3}$, Magono and Nakaamura (1965)] and graupel particles [i.e., 0.2-0.5 g cm$^{-3}$, Heymsfield (1978)]. Terminal velocities of snowflakes were calculated as in Magono and Nakamura (1965), while fall speeds of the raindrops resulting from melting were obtained from Beard (1976).

Environmental temperature and moisture profiles were based on the Hudson sounding (Fig. 2). The vertical resolution of the model runs was 50 m, while the time step used was 0.5 s. Snowflakes were released continuously at 4.45 km AGL, where they fell into the subcloud layer and after approximately 600 s produced a steady-state downdraft. The control run using these input parameters is shown in Fig. 17.

In viewing the model results, the limitations of comparing the model runs with actual observations should be noted. Since the model has an open lower boundary condition, comparisons with the Doppler analysis and aircraft measurements below about 1.5 km cannot be made. The selection of this height is based on Fig. 11, which reveals significant surface divergence extending up to this level. Caution must also be used when making comparisons between point measurements from an aircraft and radar syntheses with a numerical simulation that assumes horizontal homogeneity and a continuous supply of hydrometeors from the source. Even with these restrictions, it is believed that the model results add the needed physical insight on the forcing mechanisms of the microburst downdraft. In addition, it is believed that fully documenting the microphysical processes is of critical importance to understand the microbursts on 9 July; therefore, a model with detailed microphysics (not bulk microphysics) is essential.
Fig. 17. Plots of vertical air velocity, thermal buoyancy, relative humidity, particle surface temperature, radar reflectivity, and liquid water mixing ratio using Srivastava’s model. The numbers on the curves are the time from the initiation of the downdraft at the top. The location of the radar-observed bright band is noted. The height and the maximum downdraft speed based on the dual-Doppler analysis and the Wyoming King Air penetration are also shown for comparison. The assumed density of the snow aggregate and concentration are indicated. The diameter and melted diameter are $D$ and $D_m$, respectively.

There is good agreement between the numerical simulation of vertical air motion and data obtained from the Wyoming King Air penetration ($\approx12.8$ m s$^{-1}$) and the dual-Doppler synthesis (approximately 18.3 m s$^{-1}$, the peak gridpoint value within the microburst downdraft) in Fig. 17. This suggests that the model is capturing the primary processes that intensify the downdraft, which must include precipitation loading (which is small since the reflectivities are low) and cooling owing to phase changes of the hydrometeors. The significant negative thermal buoyancy and drying as the downdraft descends from cloud base are evident in the figure. Observational (Kessler et al. 1988; Kingsmill and Wakimoto 1991; Parsons and Kropfli 1991) and two-dimensional numerical studies (Proctor 1989) have shown that the vertical pressure gradient force below about 1 km is significant in slowing the downdraft. For the same type of storms observed in this study and with no rotation within the downdraft (the 9 July downdrafts did not exhibit any rotation), Proctor (1989) has shown that the pressure gradient force is negligible between 1.5 km and cloud base. Accordingly, it is believed that the pressure forces on 9 July were not important in the downdraft intensification between cloud base and about 0.5 km below the location of the bright band.

The observed bright band is marked on the plots of particle surface temperature and simulated radar reflectivity. As noted previously, there is a substantial delay in the melting process as the downdraft descends owing to the relatively dry environmental conditions. When the environmental temperature is 0°C, the particle surface temperature is still only $-1.76^\circ$C. Hence, it is hypothesized that sublimation cooling keeps the surface particle temperature below freezing and delays the onset of melting. In addition, there is excellent agreement between the plot of liquid water mixing ratio and the location of the observed bright band.

To examine the effects of sublimation, sensitivity experiments were run by turning off the effects of sublimation, evaporation, and melting (Fig. 18). The dashed line on the figure represents the vertical wind speed obtained in the control run. These simulations clearly indicate the importance of sublimation in providing the negative buoyancy to intensify the downdraft and are consistent with the results shown in Fig. 16a and those presented by Proctor (1989). Interestingly, negating the effects of melting and evaporation had
little effect on the downdraft intensity. It is also possible that the diminished role of evaporation and melting in driving these virga microbursts is a result of reduced precipitation mass by sublimation at the time these two processes come into play.

7. Null cases

In all of the microburst studies documented in the literature, the characteristics of the null cases (or weak downdrafts) have been largely ignored for low-reflectivity storms, in part due to the difficulty in detecting them. Understanding this type of downdraft could have important implications for nowcasting wind-shear events with a Doppler radar using attached automated detection techniques. For the latter, quantification of thresholds that distinguish characteristics of a wind-shear event from a nonevent are necessary to optimize the performance of computer algorithms. This study is strengthened by having dual-Doppler analyses that focus on the null event.

Figure 16b depicts vertical cross sections of vertical velocity, divergence, and radar reflectivity for a weak downdraft associated with storm 21 (refer to Fig. 4 for cell location). These parameters were determined using the dual-Doppler analyses discussed in section 2. Storm 21 propagated along a nearly identical storm track as storm 23a but preceded it by about 15 km. Its storm motion vector was also the same. Similar to storm 23a, this storm produced multiple downdrafts in a line oriented parallel to the storm motion vector. New downdrafts formed northeast of the older downdrafts, in proximity to the maximum reflectivity area near cloud base. Once formed, the downdrafts did not move significantly from their original position as the storm propagated northeastward.

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**Fig. 18.** Sensitivity tests on the vertical wind speed using Srivastava’s model by turning off the effects of sublimation, evaporation, and melting. The dashed gray line in the figure represents the vertical wind speed obtained in the control run.

**Fig. 19.** Vertical cross section of the horizontal area encompassed by the 10-dBZ contour for microburst storm 23a and storm 21, which produced only weak downdrafts (null case). The time when the first microburst reached the surface from storm 23a is shown in the figure. The dual-Doppler analysis times are indicated on the figure by the dashed lines. Horizontal area greater than 100 km² is shaded in gray.
The magnitudes of divergence and vertical velocity in Fig. 16b are smaller than those shown in Fig. 16a. Likewise, the magnitude of the vertical acceleration of the downdraft between cloud base and the radar bright band is considerably reduced for the null downdraft. The first downdraft produced by storm 21 was associated with a weak surface outflow that never reached microburst intensity. This is not the downdraft shown in Fig. 16b, since coordinated radar scanning commenced as this weak downdraft was reaching the surface. The remaining downdrafts in storm 21 were never able to reach the surface. Surprisingly, the radar reflectivity values in Figs. 16a and 16b are comparable for the two cases and, in fact, are higher for the null case. However, the null case has a more shallow parent cloud as determined by the vertical depth of the 15-dBZ echo above the bright band. As can be seen in Fig. 16, the 15-dBZ vertical depth decreases with time for both the microburst and null storms and is caused by the storms propagating away from the downdrafts.

After examining the PPI volume scans for a number of different parent clouds, a useful variable to consider is the horizontal area encompassed by the 10-dBZ isopleth for the entire storm. This parameter was calculated for the microburst and null case and is plotted versus time in Fig. 19. Apparent in this figure, the number (or mass) of precipitation particles associated with the microburst-producing storm is substantially larger than for the null case. The time a microburst was first detected at the surface from storm 23a is indicated in the figure. A second null case, storm 22 (see Fig. 4 for location), was also examined in this manner (not shown) and had an even smaller area than storm 21. None of the downdrafts produced by storm 22 reached the surface, unlike storm 21. These results suggest promise for the development of algorithms capable of distinguishing microburst storms from null storms in real time based on their size. However, additional work is needed to determine the statistically significant threshold between the two types of events.

Characteristics of microburst and null downdrafts within storms 23a, 21, and 22 are summarized in Table 2, as determined from the dual-Doppler analyses. A total of 12 downdrafts are examined: 6 from storm 23a, 3 from storm 21, and 3 from storm 22. Only those downdrafts that could be observed within more than one dual-Doppler volume scan were included. The maximum time interval covered for any downdraft was ±4 min from the time of maximum divergence whether at the surface within a microburst or aloft within a null downdraft. For each downdraft event, the maximum value (selected from all analysis times) for the parameters shown in the table was identified. Subsequently, the average and standard deviation of these maximum values for all downdrafts were tabulated and entered into Table 2. The average (and standard deviation) height of observance was also calculated. Key features from this table include the following.

<table>
<thead>
<tr>
<th>Parameter</th>
<th>Storm 23a</th>
<th>Storm 21</th>
<th>Storm 22</th>
</tr>
</thead>
<tbody>
<tr>
<td>Microburst-producing downdrafts</td>
<td>20</td>
<td>14</td>
<td>10</td>
</tr>
<tr>
<td>Nonmicroburst-producing downdrafts</td>
<td>14</td>
<td>14</td>
<td>16</td>
</tr>
<tr>
<td>Number of downdrafts</td>
<td>20</td>
<td>14</td>
<td>10</td>
</tr>
<tr>
<td>Maximum vertical velocity (m s⁻¹)</td>
<td>3.9 ± 3.2</td>
<td>14.2 ± 2.6</td>
<td>12.2 ± 1.4</td>
</tr>
<tr>
<td>Maximum divergence (10⁻³ s⁻¹)</td>
<td>14 ± 1.8</td>
<td>14 ± 1.6</td>
<td>8 ± 1.0</td>
</tr>
<tr>
<td>Maximum downdraft (km AGL)</td>
<td>6.0 ± 0.6</td>
<td>6.0 ± 0.8</td>
<td>1.0 ± 0.6</td>
</tr>
<tr>
<td>Maximum reflectivity (dBZ)</td>
<td>≥ 15</td>
<td>≥ 15</td>
<td>≥ 15</td>
</tr>
<tr>
<td>Maximum vertical depth of downdraft (km)</td>
<td>≥ 3.6</td>
<td>≥ 2.1</td>
<td>≥ 1.7</td>
</tr>
</tbody>
</table>

*The reflectivity values were "linearized" before summation.
**Indicates reflectivity for three downdrafts had no bright band.
1) Microburst downdrafts are wider and are associated with a more intense bright band than the null cases. Half of the null cases had no bright band. The difference in brightband intensity can be attributed to the greater mass flux of particles through the melting level.

2) Consistent with individual case studies presented in Fig. 16, the maximum radar reflectivity of the parent clouds does not vary significantly between the two types of downdrafts. However, the vertical depth of the reflectivity greater than 15 dBZ is over 1 km greater for the microburst storms.

3) The maximum subcloud convergence is only slightly higher for the microburst downdrafts. This suggests that this parameter could be a misleading precursor for predicting microbursts. Additional information used in conjunction with the subcloud convergence, such as the area of the 10-dBZ isopleth, may be necessary to differentiate microburst-producing storms from null events.

8. Summary and discussion

An analysis of a storm that produced a series of low-reflectivity microbursts was presented. These types of microbursts are particularly hazardous to aircraft since the parent cloud and pendant virga shafts appear innocuous. The microburst downdrafts from storm 23a develop at the location where the virga shafts are, visually, the lowest and opaque. However, as the downdraft intensifies, sublimation and (to a smaller extent) evaporation rapidly deplete the hydrometeors. The latter conclusion was derived from radar brightband observations combined with numerical simulations with a one-dimensional microphysical model. The conclusion that sublimation of snowflake aggregates is the primary forcing mechanism for this type of microburst is consistent with the hypothesis advanced by Proctor (1989).

As the downdraft reaches the surface, the axis of maximum negative vertical velocities subsequently produces a relative low-reflectivity region and a transparent region of the virga shafts. Radar reflectivity data combined with cloud pictures revealed the location of the radar bright band to be at the visible termination of the virga shafts (i.e., the melting level). This appears to be the first conclusive evidence that virga shafts are ice particles rather than raindrops, confirming recent speculations by Frazier and Bohren (1992).

A schematic model summarizing this case is shown in Fig. 20. The series of downdrafts in various stages of development are indicated. The visual character of the virga from relatively low pendant shafts as the downdraft first initiates to the development of a virga hole as the microburst achieves its most intense outflow stage is also shown.

Possible implications for nowcasting these wind-shear events with a single-Doppler radar require the identification of a pronounced bright band in the formidable stages of the microburst and a relatively large horizontal area encompassed by the 10-dBZ contour. In the present case this area had to exceed between 70 and 100 km² at levels between 3 and 6 km before microbursts reached the surface. However, the results presented in this paper suggest that an integration of detailed Doppler radar information and visual characteristics of the clouds is necessary in order to produce the most effective nowcast.

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