FORCING MECHANISMS OF THUNDERSTORM DOWNDRAFTS

by

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ABSTRACT

The role of various forcing mechanisms of thunderstorm downdrafts has been explored. Four case studies of downdrafts are presented. Radar data from five closely spaced radars are used to conduct direct triple Doppler calculations of the full three-dimensional wind field in three of the cases; data from three closely spaced radars are used in the remaining case. Differential reflectivity measurements are used to diagnose the presence of ice in the downdraft regions. Data from a dense array of surface mesonet stations are used to diagnose the thermodynamic characteristics and origin levels of outflow air from the downdrafts. A one-dimensional parcel-following model, which incorporates evaporation, melting and entrainment, and which is strongly constrained by detailed radar and surface observations, is used to diagnose the relative importance of various forcing mechanisms.

The studied downdrafts, while originating at different altitudes, ranging from 2 km to 4.5 km, are found to be forced by the same basic mechanisms: cooling due to the evaporation of precipitation inside the cell and precipitation loading. The deeper downdrafts are nearly neutrally thermally buoyant above 2 km and are driven downwards by precipitation loading. The evaporation of precipitation is crucial to the maintenance of neutral buoyancy. The intensity of the modelled downdrafts is insensitive to whether the entrainment source is cloudy or outside environmental air. This is because the positive buoyancy of entrained cloudy air is offset by the supply of rapidly evaporating cloud droplets into the downdraft. Cloud droplet evaporation is found to contribute about 40% of the total evaporation if entrainment occurs from primarily inside the cloud. In the two downdrafts with deep origins, near 4.5 km AGL, engulfment of outside air into the upper levels of the downdrafts is observed.

Ice phase precipitation is found to be unimportant in the forcing of these downdrafts. A very narrow, less than 1 km horizontal scale, region of mixed phase precipitation is present in one of the downdrafts at one observation time. The melting of ice is found to be of secondary importance compared to other forcing mechanisms. Ice phase precipitation is not observed in the three remaining downdrafts.

Three of the downdrafts, exhibiting different origin levels and intensities, occur simultaneously within a 10 km region illustrating the variety of events that can develop in virtually identical environments. The studied downdrafts occur in moist, low-cloud base environments which contrast with those common in the High Plains and western USA, where the dominant forcing mechanisms probably differ.

Thesis Supervisor: Dr. Earle Williams
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CHAPTER 1: Introduction

A. Brief Overview

Thunderstorms produce downdrafts that have widely varying structures and surface thermodynamic characteristics. Despite these differences, many of the downdrafts have very similar forcing mechanisms. The varied structures and similar forcing mechanisms of four downdrafts are documented in this thesis through the use of an unprecedentedly dense radar network and surface mesonet array, in conjunction with a simple one-dimensional model. It is shown that, despite varied origin levels and peak downward velocities, the dominant causes of air parcel acceleration are precipitation mass loading and in-cloud evaporation of raindrops. Sub-cloud evaporation of raindrops is not found to be significant. Evaporation of cloud droplets is found to be only of secondary importance. Melting of ice phase precipitation is not observed in three of the downdrafts while in the fourth, the evidence for ice is equivocal and the role of melting ice as a forcing mechanism is found to be of secondary importance.

The three-dimensional evolution of the downdrafts is examined by conducting analyses of radar data from a five-radar network near Huntsville, Alabama, in 1986 and from a three-radar network that has been near Orlando, Florida, since 1990. The geometry of these networks is such that highly accurate three-dimensional wind fields can be synthesized throughout much of the depth of the downdraft regions.

These fields reveal that both deep (greater than 4 km vertical extent) and shallow (about 2 km vertical extent) downdrafts can exist simultaneously in very similar environments. The downdrafts form in the high (greater than 45 dBZ*) reflectivity regions of the thunderstorm cells, though not necessarily in the center of the highest reflectivity cores. Differential reflectivity measurements indicate that ice phase precipitation is not predominant in the downdraft regions of any of the cells.

The thermodynamic characteristics of the outflow air from the downdrafts is examined. In Huntsville, there was a dense network of 71 surface mesonet stations, spaced at roughly 2-3 km intervals, providing minute-by-minute data; in Orlando, a more limited mesonet

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* Radar reflectivity, measured in logarithmic units of dBZ, is defined and discussed in Chapter 2.
array was in operation. This data, when combined with the radar observations and model simulations, confirms the varied origin levels and similar forcing mechanisms active in the studied downdrafts.

B. Possible Forcing Mechanisms

Downdrafts in thunderstorms have been observed for many years; the early work of Byers and Braham (1949) is notable. Somehow, air that has been moving upwards in the thunderstorms, or air from the outside environment, that, presumably, has little initial vertical motion, is accelerated towards the ground.

There are a multitude of possible mechanisms may be invoked to accelerate air parcels downwards. The most plausible mechanisms, the existence of which are supported by observational or modelling evidence, can be profitably separated into four categories. These are:

1. Downwards buoyancy caused by cooling from the melting of ice in the cloud,
2. Downwards buoyancy caused by cooling from the evaporation of precipitation into sub-saturated air that has been entrained from outside of the cloud,
3. Downwards buoyancy caused by cooling from the evaporation of raindrops into sub-saturated air that exists below cloud base,
4. Downwards buoyancy caused by the weight of precipitation particles in the cloud.

While it is not clear that any of these processes, shown schematically in Figure 1.1, act exclusively, it is useful to examine the qualities of downdrafts (as manifested by surface observations of thermodynamic characteristics and outflow strength) that would result from each of these mechanisms acting individually. These are not presented as plausible scenarios of downdraft forcing but, rather, serve to provide distinctions in the possible observable qualities of downdrafts forced by these varied mechanisms.

1. Melting of Ice

If ice particles fall into an air parcel and melt, the temperature of the air parcel will drop and its buoyancy will decrease. The magnitude of this drop is dependent on the mass of ice that melts and the mass of air that is affected by the cooling. The air will cool according to the following equation:
\[ T = T_{\text{initial}} - \frac{L_f M_i}{C_p M_a}, \]

where \( L_f \) is the latent heat of fusion, \( M_i \) is the mass of ice that melts, \( C_p \) is the heat capacity of air at constant pressure and \( M_a \) is the mass of air affected. If \( M_a = 765 \text{ kg/1000 m}^3 \), consistent with a 1 km thick melting layer, and \( M_i = 6 \text{ kg/1000 m}^3 \), consistent with adiabatically lifted liquid water contents (Musil and Smith, 1989 found liquid water contents of up to 14 g/m\(^3\) in Huntsville storms.), and a constant flux of ice into the melting layer is assumed, the rate of cooling is found to be 1 K - 2 K / minute. This would produce downward accelerations of \( a = \frac{g \Delta T}{T} = -0.15 \text{ m/s}^2 \), where \( \Delta T \) is the cooling during an assumed 5 m/s, three minute journey through the melting layer. Since this acceleration is four times larger than any observed in the downdrafts studied here, this mechanism alone could explain the observed downward motions.

The value of equivalent potential temperature, \( \theta_e \), is nearly conserved in an air parcel during evaporation and condensation. This conservation is violated during processes involving the ice phase of water.* In the example presented above, by producing coolings of approximately 5 K at the base of the melting layer (assuming cooling rates of 1-2 K for 3 minutes), melting could cause \( \theta_e \) drops on the order of 10 K (assuming \( d\theta_e/dT \) is about 3 K/K). These depressed values might be observable at the ground.

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* Equivalent potential temperature, \( \theta_e \), is used extensively throughout this thesis. As defined by Bolton (1980), "The equivalent potential temperature will be taken... to be the final temperature, \( \theta_e \), which a parcel of air attains when it is lifted dry adiabatically to its lifted condensation level, then pseudo-wet adiabatically (with respect to water saturation) to a great height (dropping out condensed water as it is formed), then finally brought down dry adiabatically to 1000 mb." The value of \( \theta_e \) in an air parcel is conserved during evaporation and is useful for tracing downdraft air parcel origins. An air parcel's value of \( \theta_e \) is not conserved if it mixes with air parcels exhibiting different values of \( \theta_e \), if ice melts in it, or if it experiences other diabatic heating or cooling. The formulation of Bolton (1980) is used in this study:

\[
\theta_e = T \left( \frac{1000}{P} \right)^{0.2854(1-0.00028r)} \exp \left[ \frac{3.376}{\left( \frac{1}{T_a-56} + \frac{\ln(T/T_d)}{800} \right)} + 0.00254 \right],
\]

where \( T \) is temperature (K), \( T_d \) is dewpoint temperature (K), \( P \) is pressure (mb), and \( r \) is water vapor mixing ratio. \( \frac{d\theta_e}{dt} \) at 1000 mb = 4 K/K, \( \frac{d\theta_e}{dR} \) at 003 K = 1 K/%, \( \frac{d\theta_e}{dt} \) at 000 mb = 600 K/K, \( \frac{d\theta_e}{dR} \) at 273 K = 0.4 K/%.
A downdraft forced by the melting of ice would probably have origins in the melting layer. Large amounts of ice phase precipitation, moving downward, would have to be present in order to provide the melting ice flux. The possible existence of deep downdrafts, downward precipitation velocities in the melting layer, and the presence of ice could be detected with radar data available from Huntsville. Differential reflectivity data, which is used to distinguish liquid from solid phase precipitation, was not taken in Orlando, so the presence of ice could not be explicitly determined there. Depressed values of $\theta_e$ associated with the melting of ice might be observed at the ground.

Melting and sublimation processes are the only short time-scale sinks of $\theta_e$. Unequivocal evidence of the role of melting ice would exist if surface observations of $\theta_e$ in an outflow could be found in which the values of $\theta_e$ were less than the minimum values observed in the three-dimensional environment of the storm. In this case, no possible mix of mid-level and low-level air, saturated or not, could produce the observed values. The dataset from Huntsville provided an excellent opportunity to attempt to find such a case. Thirty-seven outflows impinged on mesonet stations on days when proximity soundings were taken at the Redstone Arsenal, just to the southeast of the mesonet array (the southern portion of the array is shown in Figure R.6). Figure 1.4 (bottom panel) shows the difference between the mid-level minimum $\theta_e$ and the minimum surface $\theta_e$ during outflows, $\Delta \theta_e$. $\theta_e$ at the surface was never lower than the mid-level minimum. Two soundings, one that was typical of many soundings, though exhibiting a relatively small surface $\theta_e$ drop, and one that shows the largest observed $\Delta \theta_e$, are shown in Figure 1.3. The 03 June 1986, high $\Delta \theta_e$ case provides strong, though not unequivocal, evidence of the role of ice. Either almost pure (at least 82%) 675 mb environmental air was brought from mid-levels to the mesonet station, or melting ice contributed to the depressed values of $\theta_e$ observed at the surface.

These discussions ignore the influence of entrainment on the descending downdraft parcels. Entrainment would dilute the $\theta_e$ depressed parcels with high $\theta_e$ cloudy air or low $\theta_e$ environmental air. Entrainment of environmental air would also provide opportunity for continued cooling due to the evaporation of raindrops, which is discussed below. Entrainment of cloudy air would provide an opportunity for continued cooling due to evaporation of entrained cloud droplets. Adiabatic compression would cause the parcels to warm and become sub-saturated which would also provide an opportunity for further
evaporation of either cloud droplets or raindrops. So, melting of ice could occur without being detected by surface observations.

2. Saturation of Entrained mid-level air

If sub-saturated environmental air is engulfed into a storm at middle or high levels, evaporation will occur as the air either mixes with air laden with cloud droplets (which will immediately evaporate) or as raindrops fall through it (and evaporate slowly). The amount of cooling that would occur depends on the dryness of the engulfed environmental air. Maximal cooling will occur if environmental air enters the storm in pure (unmixed) form, saturates as rain falls into it, and cools to its wet bulb temperature. On 11 July 1986, a typical summer day in Huntsville, Alabama, the wet bulb depression at 4 km altitude was about 4 K (see Figure 3.B.2). Therefore downward accelerations of  

$$a = \frac{g\Delta T}{T} = -0.15 \text{ m/s}^2$$  

could result. As was the case with melting ice, this is four times larger than the accelerations observed in this study. The value of $\theta_e$ in the downdraft would initially be low, equal to the environmental values at the engulfment level, since the proposed evaporative mechanism conserves $\theta_e$. The values of $\theta_e$ at mid-levels (near 4 km altitude) were typically about 20 K lower than the surface, as shown in the bottom panel of Figure 1.4. $\theta_e$ drops of 20 K would be easily observable at the ground. As with the discussion of melting ice, the role of continued entrainment during descent is neglected in these simple calculations. Entrainment would dilute the value of $\theta_e$ in the descending parcels and complicate the interpretation of observations at the surface.

A downdraft caused by the cooling of entrained mid-level low $\theta_e$ air would be likely to have origins at these mid-levels. Horizontal convergence near the cloud edge might be detectable since the engulfed parcels would probably mix with cloudy air and cloud droplets would immediately evaporate and cool the parcels which would accelerate downwards due to negative buoyancy forces. In addition, raindrop evaporation would begin to cool the sub-saturated parcels immediately.

The availability of dry mid-level air would be expected to enhance the amount of cooling that the engulfed air parcels experience and enhance downdraft accelerations. Data presented in Figure 1.2 shows that there is no systematic correlation between outflow strength and $\Delta \theta_e$, previously defined. Strong outflows, with differential radial velocities of over 20 m/s, exhibit the complete range of $\Delta \theta_e$. Of course, $\theta_e$ is not a buoyancy
variable, though it tends to be well correlated with buoyancy since dry environments usually exhibit low $\theta_e$ and low virtual temperature, $T_v$. Figure 1.5 shows that there is also no correlation between 700 mb or 850 mb dewpoint depressions (representing the potential for evaporative cooling in engulfed parcels) and outflow strength.

Both the melting of ice and the entrainment of low $\theta_e$ mid-level air would result in observations of low $\theta_e$ at the ground. With the possible exception of unusual cases like 03 June 1986, it may be impossible to distinguish between these mechanisms with surface thermodynamic data alone. Fortunately, differential reflectivity data (see Chapter 2) were available in Huntsville and will be used to diagnose the possible presence of ice.

3. Cooling of Sub-saturated Sub-cloud Air

If raindrops fall into a sub-saturated sub-cloud environment, they will evaporate and cool the sub-cloud air. The amount of cooling that will occur depends on the dryness of the sub-cloud air. While sub-cloud environments can be very dry in the western United States, they are usually fairly moist in the East. Cloud bases in the East are typically much lower than those observed in the West, typically only about 1 km above ground level. A crude assumption is that air below cloud base will be cooled to its wet bulb temperature by raindrop evaporation. Typical sub-cloud wet bulb depressions in Huntsville are on the order of 4 K and, thus, accelerations of $a = \frac{g \Delta T}{T} = -0.15 \text{ m/s}^2$ could result. This is similar in magnitude to the accelerations that resulted in the simple melting and engulfment calculations and is significantly larger than the highest accelerations observed in this study. Values of $\theta_e$ in the downdraft would be equal to the sub-cloud environmental values which are typically near (within 5 K) the surface values as can be seen in the two soundings presented in Figure 1.3.

Downdrafts forced by sub-cloud evaporation of raindrops would have origins near cloud base. Horizontal convergence near cloud base would be measurable by radar. Surface

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*Virtual temperature, $T_v$, following Pruppacher and Klett (1980) "... is the temperature which dry air would have to have in order for its density to match that of the actual air." This quantity is used to compare air parcel densities when conducting buoyancy calculations. $T_v$ is calculated as

$$T_v = \frac{T}{1-(1-.622)*e/P}$$

where $e$ is the partial pressure of water vapor (mb), $T$ is air temperature (K), and $P$ is air pressure (mb).
mesonet observations would show only small $\theta_e$ drops (less than 5 K). Drops of this magnitude are frequent in Huntsville and dominant in Denver, Colorado (see Figure 1.4, top and middle panels). The relationship between downdraft strength and sub-cloud dryness is more problematic. It might be expected that drier sub-cloud environments would be conducive to stronger downdrafts through their greater evaporative cooling potential and greater depth. But, as noted in Srivastava (1985), dry sub-cloud environments may actually suppress downdrafts because of their low $T_v$. The relationship between surface relative humidity and outflow strength in Huntsville and Denver are shown in Figures 1.6 and 1.7. (Denver data is from Isaminger, pers. comm. and DiStefano, 1988.) The relative wetness of the surface environment in Huntsville is evident. There is almost no correlation between the sub-cloud dryness and the strength of the outflows. This is in contrast to the findings of Srivastava (1985) who predicted a negative correlation. A possible reason for this is that Srivastava (1985) did not consider that surface relative humidity and cloud base height are probably negatively correlated. Thus, downdraft air parcels in a dry sub-cloud environment, with its resultant high cloud base, will have a greater distance in which to accelerate.

4. Precipitation Loading

Water mass inside clouds necessarily exerts a drag force equal to its weight on the air parcels through which it passes. This force depends on the precipitation and cloud water contents in the cloud. If $\rho_L = 6 \text{ g/m}^3$ precipitation water content and $\rho_v = 1 \text{ g/m}^3$ cloud water content are assumed, (Musil and Smith, 1989), then accelerations of

$$a = \frac{g(\rho_L + \rho_v)}{\rho_a} = .1 \text{ m/s}^2$$

result. This is somewhat smaller than the maximum accelerations that are predicted for the other mechanisms, but is still more than sufficient to explain the accelerations observed in this study. Values of $\theta_e$ in the precipitation loaded downdraft would be equal to the in-cloud values.

Downdrafts forced by precipitation loading could originate at any level where high precipitation and cloud water contents exist (providing that some process has increased local water contents above the levels produced by reversible ascent). They would exist primarily in the high reflectivity regions of the cloud where the forcing due to the loading would exceed any possible forcing due to upwards thermal buoyancy. Since the amount of forcing due to this drag is primarily dependent on the precipitation water content (Cloud
water contents are typically near 1 g/m³ and rarely exceed 3 g/m³*), a correlation between maximum storm reflectivity and outflow strength might be expected. Figures 1.8 and 1.9 show this relation for Huntsville and Denver. All the outflow events in Huntsville occurred in storms that exhibited maximum reflectivities of at least 45 dBZ. There was a moderate correlation with reflectivity and the strongest events occurred in storms that exhibited reflectivities of 60 dBZ or more. In Denver, there was almost no systematic correlation between outflow strength and maximum storm reflectivity. (This is consistent with the results of Wilson et al. (1984) who presented data taken in Denver in 1982.) In contrast to Huntsville, there were outflows from storms that contained only low maximum reflectivity. It is notable, however, that all the outflows which exhibited values of ΔV greater than 26 m/s, originated in storms that contained maximum reflectivities of at least 50 dBZ. Observed values of θₑ at the surface under precipitation loaded downdrafts would depend on the origin levels of the downdrafts. Shallow downdrafts that are confined to the lowest regions of the storm would exhibit high values of θₑ, while downdrafts originating at mid-levels would contain the lower θₑ air that is typical of the mid-level regions of clouds (Musil and Smith, 1989, Byers and Braham, 1949).

As with the discussion of melting ice and engulfment, the role of continued entrainment during descent is neglected in these simple calculations.

Any or all of these mechanisms are sufficient to force downward accelerations of the magnitude that are observed. It is impossible, on the basis of the above simple calculations and statistical correlations, to conclude much about what mechanisms are dominant. The only statistically significant correlation that exists is between maximum reflectivity and outflow strength in Huntsville. While this may point toward precipitation loading as a dominant mechanism, higher precipitation water contents would also allow greater cooling due to melting and more rapid cooling due to evaporation.

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* Musil and Smith (1989), Waldvogel et al. (1987), Musil et al. (1986b), Krauss and Marwitz (1984), Heymsfield and Musil (1982), Musil et al. (1978), Sand (1976) Musil et al. (1974), though Paluch (1979) observed higher values in non-precipitating cumuli and Musil et al. (1986a) found values over 6 g/m³ in the updraft (40 m/s) of a High Plains supercell thunderstorm.
C. Entrainment

The nature and effect of entrainment into downdrafts has been ignored in the simple calculations presented above. Laboratory models of descending parcels entrain as similarity plumes (Morton et al., 1956). These plumes entrain air from their proximate environment as they descend. There is disagreement as to the nature of the mixing that occurs in clouds. Paluch (1979) shows that, in some non-precipitating cumuli, the interior air parcels are composed of varying mixes of cloud base and cloud top air. Thus little or no air is inferred to be entraining laterally from the edges of the cloud. These results have been confirmed in non-precipitating cumuli by Jensen et al. (1985), among others. Emanuel (1981) develops a penetrative downdraft model and simulates deep downdrafts, entraining only in-cloud air. Raymond and Blyth (1986) present modeling results and aircraft observations. The observations of detrainment fluxes are discussed in more detail in Raymond and Wilkening (1985).) suggesting that detrainment from updrafts in primarily non-precipitating cumuli occurs at several levels in the clouds. Observations of Blyth et al. (1988) and Raga et al. (1990), also in cumuli, suggest that entrainment is occurring at many levels in the clouds. Both studies present evidence that in-cloud parcels consist of mixes that include air from at or near the aircraft penetration levels, implying that lateral entrainment has occurred. These observations are consistent with the crude observations of Byers and Braham (1949) that showed balloons in the out-of-cloud environment being drawn (possibly) towards the thunderstorm cell. Extrapolation of any of the observations of cumuli to large precipitating thunderstorms is difficult to justify. Therefore, the true nature of the entrainment process in thunderstorms remains unknown.

The observations that are presented in this study show that, at least in some downdrafts, entrainment from outside the cells is occurring, consistent with the observations of Byers and Braham (1949). In the model simulations that are presented, the results are not extremely sensitive to the type of entrainment that is specified. This is because entrained in-cloud air, while positively buoyant, supplies cloud droplets that evaporate in the downdraft and allows descending parcels to remain cool relative to the environment. Entrained out-of-cloud air, while cool and dry, contains no cloud water. Therefore, downdrafts that entrain this type of air tend to dry and warm as they descend.

By continuously entraining either in-cloud or out-of-cloud air, downdrafts could become diluted. This would impact on the thermodynamic observations at the surface and
complicate the diagnoses of the source regions of outflow air since no single source region would exist.

D. Previous Modeling and Observational Results

The basic structure of the windfields and physics of thunderstorms has been examined for many decades. The early work of Byers and Braham (1949) presents a mature phase thunderstorm with a downdraft extending from about 25,000 ft or 7.6 km to the ground. They hypothesized that precipitation drag initiated downward motion as water loading increased and thermal buoyancy decreased due to entrainment of environmental air during the mature phase of the storm. The downdrafts observed in this study, originating at 5 km or below, were not nearly as deep as the ones described in Byers and Braham (1949). Byers and Braham (1949) indicate that downdrafts originate in regions containing ice-phase precipitation while the observations presented here indicate that liquid-phase precipitation is predominant in the upper regions of the downdrafts.

More recently, there have been many additional studies of thunderstorms and downdrafts. The intent here is not to present an exhaustive review of all the available literature on thunderstorms, but, rather, to discuss selected studies that bear directly on forcing mechanisms and radar observations.

The entrainment of sub-saturated air into a cloud causing evaporative cooling has been discussed by Emanuel (1981). He details a downdraft that is a similarity plume, entraining cloudy air during its descent and cooling due to evaporation. While much of his paper deals with cloud-top entrainment events, the possibility of mid-level entrainment in cumulonimbi, consistent with the inflow that is observed in this study, is discussed. In the simulations of plumes entraining ice crystals, the issue of slow evaporation is discussed. This is very relevant to the precipitating clouds observed in this study, since raindrops evaporate very slowly.

Srivastava (1985) presents a one-dimensional model in which evaporation in a deep sub-cloud layer is simulated. He finds that the strength of the simulated downdrafts is dependent on, among other things, the sub-cloud relative humidity (higher humidity = stronger downdraft because of virtual temperature effects) and rainwater mixing ratio (higher mixing ratio = stronger downdrafts). Observations during 1987 in Denver, see Figure 1.7, indicate that there was, in actuality, very little correlation between sub-cloud
relative humidity and surface outflow strength (which, presumably, is well correlated with downdraft strength). The dependence on rainwater mixing ratio, as inferred from maximum storm reflectivity is also not observed in Denver, see Figure 1.9. Srivastava (1985) presents data that shows that there is a weak inverse dependence between storm reflectivity and environmental lapse rate. Since the highest reflectivity storms were forming in environments that were the least favorable to the descent of downdraft parcels, perhaps this accounts for the lack of correlation between maximum storm reflectivity and outflow strength.

As noted in Srivastava (1985), thunderstorms in moist environments may differ significantly from the ones that he attempted to simulate. Observations in Byers and Braham (1949) for Ohio and Florida and those presented here for Huntsville and Orlando, do not exhibit the extreme sub-cloud dryness of his simulated downdrafts. The outflows and thunderstorm environments in Huntsville and Denver differ in the following four ways:

1. The pre-event surface relative humidities are higher in Huntsville, implying that sub-cloud evaporative processes will be much slower. (Evaporation of raindrops into air at 20% relative humidity is twice as fast as evaporation into air at 60% relative humidity, see Appendix 2.)

2. Outflows in Huntsville only occur under storms containing high maximum reflectivities, generally over 45 dBZ, see Figure 1.8.

3. Outflows in Huntsville exhibit considerably larger $\theta_e$ drops than those in Denver. The mean $\theta_e$ drop in Huntsville is near 10 K, while the mean $\theta_e$ drop in Denver is nearly 0 K, see Figure 1.4, top and middle panels.

4. The distance over which sub-cloud forcing mechanisms can act is severely limited in Huntsville due to the small distance between the low cloud bases and the surface outflow layer, as discussed in Chapter 5.

These observations suggest that precipitation loading may be playing a more significant forcing role, relative to sub-cloud evaporation, in the moister environments typical of the eastern United States.
The case study observations presented in Chapter 3 and Chapter 4 will show downdrafts that are clearly dissimilar to those described in Srivastava (1985) in that they originate at 2 km - 5 km above the surface in high reflectivity and are forced by a combination of precipitation loading and in-cloud evaporation. All the studied downdrafts originate well above cloud base, which is near 1 km above the surface.

Several workers have explored the role of ice in forcing downdrafts with both models and observations. Srivastava (1987) used a model which incorporated the effects of evaporation, melting and precipitation loading. He simulated downdrafts in storms with high cloud bases, typically 3.8 km AGL, and found that the inclusion of melting ice could significantly increase peak downdraft velocities. The high cloud bases in his simulated storms were typical for Denver, but very different than in the moist environments typical in the East.

Proctor (1989) studied the sensitivity of peak downdraft strength to various changes in environmental soundings and to precipitation phase and amount. The downdrafts strength was very sensitive to variations in environmental soundings. (This is consistent with the model sensitivity results presented here in Appendix 2.) The method used to test this sensitivity probably tested his model and did not elucidate any real physics. This is because the sensitivity was tested by placing identical clouds in each environment and then calculating the peak downdraft velocity. Since it is likely that different environments would produce cells with dissimilar characteristics, this is not a valid test of environmental sensitivity. His results also show that the peak simulated downdraft velocities were strongly dependent on precipitation phase. Precipitation in the form of snow produced the strongest downdrafts, rain the weakest, in dry environments. In wet environments, hail produced the strongest downdrafts and snow the weakest, though the variation was small.

Case studies of particular storms have discussed possible forcing mechanisms of downdrafts. Tuttle et al. (1989) present observations of a Huntsville storm (20 July 1986) that occurred well to the east of the surface mesonet. By using differential reflectivity, ZDR, measurements, discussed here in Chapter 2, they determined that there was significant ice phase precipitation down to low levels in the storm. They conclude that precipitation loading, melting of ice phase precipitation, and evaporation of raindrops are all important in forcing the observed surface outflow. The vertical structure of the downdraft is not presented, only the storm averaged vertical air velocities. These average velocities are negative below 2 km and above 8 km. To estimate the role of the various forcing terms
it would be useful to know the entire structure of the downdraft, particularly near its origin level.

Kingsmill and Wakimoto (1991) analyse this structure and show that the downdraft that causes the surface outflow originates at 4 km in the high reflectivity regions of the cell. While the downdraft originates at 4 km, it "intensifies almost exclusively within the sub-cloud layer (0-2 km)". They also note a reflectivity notch at mid-levels that is associated with a downdraft that does not reach the surface.

Differential reflectivity analyses of the same storm are presented in Wakimoto and Bringi (1988). They show a region of low differential reflectivity (below 1 dB), indicative of ice phase precipitation, in the high reflectivity downdraft region. Similar analyses for two other downdrafts (13 July 1986 and 16 July 1986) are also presented. These indicate that melting ice may play a significant role in many Huntsville downdrafts.

These observations contrast somewhat with those that will be presented in this study. Two of the downdrafts studied here show intensification above 2 km, well above cloud base. A reflectivity notch is observed here in cell S, but in this case it is associated with the outflow producing downdraft. No feature corresponding to the non-ground-reaching downdraft is observed in the case studies presented here. In addition, the melting of ice is not found to be significant in this study.

Kingsmill and Wakimoto (1991) conclude that precipitation loading is initially the dominant forcing mechanism, but that negative thermal buoyancy observed below cloud base eventually becomes the dominant forcing mechanism. As will be discussed in Chapter 3, even when thermal buoyancy is not negative, evaporation of raindrops may be playing a crucial role in maintaining near neutral buoyancy in parcels that can then be dragged downwards by precipitation loading.

Wolfson (1990) presents an analyses of a downdraft that occurred near Memphis, Tennessee on 25 June 1985. She presents evidence of an outflow that exhibits very low values of $\theta_e$, very near the environmental minimum. Dual Doppler radar analyses indicates a notch in the reflectivity field that is coincident with the low level downdraft. (Vertical velocities could not be calculated above 1.5 km.) A cloud edge entrainment is inferred from the notching of reflectivity and a 4 km origin level is inferred by comparing the outflow $\theta_e$ with the environmental sounding.
Mahoney and Elmore (1991) study two surface outflows that originate from a single storm that formed near Denver, CO. The associated downdrafts differ significantly in structure. Radar and surface thermodynamic evidence is presented. The first event, MB1, is described as the downward side of a horizontal rotor. Ensemble average trajectories are presented that indicate that the air at the surface originated 2 km to 3 km above the surface, which was consistent with the radar observations. Unfortunately, no surface thermodynamic evidence is presented for MB1. Since some of the presented trajectories originate in the sub-cloud, high $\theta_e$, layer, observations in the outflow that indicate no $\theta_e$ drop would confirm the shallow origin of the downdraft air. Surface observations under MB2, the second event, are presented; the value of $\theta_e$ drops during the outflow. The lowest values are equal to values observed in the environment 2 km to 3 km above the surface, suggesting that as the origin level of the observed surface air. The presented radar and trajectory analyses indicate that the downdraft originated at levels above 3 km AGL. While the presented radar evidence shows a downdraft extending to over 5 km AGL, the vertical velocities (calculated by integrating the mass continuity equation upwards from the ground with no upper-level boundary condition) may be unreliable at that level. Some of the results of the Mahoney and Elmore (1991) study parallel those of this thesis. As the authors note, downdrafts with very different structures can form in similar environments. The authors attempt to use surface mesonet observations to aid in the analyses but this is difficult due to the sparse nature of the network; typical horizontal spacings were near 10 km as opposed to 2 km in Huntsville and 4 km in Orlando. Since dual Doppler techniques are used to determine vertical winds, and since the cloud tops were not sampled, mid-level winds were difficult to determine accurately.
E. Motivation

Many studies have explored the nature of thunderstorm downdrafts but significant questions still remain. Modeling studies reveal that many factors can affect downdraft strength and structure; therefore these studies do not shed much light on which factors are, in actuality, most important. Statistical analyses fail to reveal the relative importance of various mechanisms. It is therefore necessary to conduct detailed case studies in which the structures of downdrafts are analyzed in order to correlate modeling studies and statistical results with reality. To date, most detailed studies of individual storms have involved Western or High Plains events. (The 20 July 1986 case in Huntsville, VA is the notable exception.) The environment and probably the dominant forcing mechanisms of downdrafts are very different in the East. Documentation and analyses of Eastern downdrafts and forcing mechanisms are needed.

Two factors motivated this particular study:

1. There is a need for detailed case studies of the wet environment downdrafts common in the Eastern United States in order to better understand their structure and associated physics. Studies that can resolve the small scale features and rapid evolution of these downdrafts would be particularly useful.

2. Uniquely useful datasets are available from Huntsville and Orlando.
   a. Huntsville:
      1. A dense network of five Doppler radars which are arranged so near each other that direct retrieval of the three-dimensional particle motion field without resort to integration of the equation of mass continuity is possible
      2. A dense network of 71 surface mesonet stations with mean inter-station gaps of 2 km - 3 km which permits detailed analyses of the thermodynamic qualities of the outflow air in comparison to the radar detected downdrafts
      3. A polarization diversity radar that measures differential reflectivity so that the phase of precipitation particles can be deduced
   b. Orlando:
      1. A dense network of three Doppler radars which are arranged so near each other that direct retrieval of the three-dimensional particle motion field without resort to integration of the equation of mass continuity is possible
      2. A dense network of about 30 surface mesonet stations with mean inter-station gaps of 4 km which permits detailed analyses of the thermodynamic qualities of the outflow air in comparison to the radar detected downdrafts
F. Organization of this Thesis

• Chapter 1 contains the introduction and review of past research that has just been read.

• Chapter 2 discusses the radar analyses techniques used in this study including triple Doppler techniques and errors, methods of integration, gridding, precipitation terminal velocity and precipitation water content determinations and differential reflectivity measurements.

• Chapter 3 presents case studies of three downdrafts that occurred in Huntsville on 11 July 1986 including radar and surface mesonet analyses and modelling results.

• Chapter 4 presents a case study of a downdraft that occurred in Orlando on 03 October 1990 including radar and surface mesonet analyses and modelling results.

• Chapter 5 summarizes the results of the case studies.

• Appendix 1 presents an explanation of the $\theta_e$ jumps observed at several mesonet stations.

• Appendix 2 presents a description of the one-dimensional model used in Chapter 3 and Chapter 4.

• Appendix 3 explains the adjustments that have been made to some mesonet data to correct for wetting of the humidity elements during rain.

• Appendix 4 discusses the relative merits of dual Doppler and triple Doppler methods for calculating wind fields and includes empirical comparisons.

Figures are labelled according to the Chapter and Section of the associated text (2.B.# in Chapter 2, Section B, for example). The exception is the figures describing the individual downdrafts in Chapter 3, Section D, which are labelled according to the downdraft to which they apply (R.13, describes cell R, for example).

For a quick review of the case study data and results, the reader is referred to Figures R.13-R.16 (Chapter 3.D.1.ii), F.14-F.19 (Chapter 3.D.2.iv), S.14,S.17, S.19-21, S.21b (Chapter 3.D.3.iv), and 4.D.4b,4.D.7,4.D.8,4.D.10 (Chapter 4.D), and Table 5.1. Those figures will show the varied structures of the downdrafts and Table 5.1 presents a comparison of the forcing mechanisms and structures of the downdrafts studied here as well as three others studied elsewhere.
G. Note on terminology

Frequent use of the terms in-cloud, out-of-cloud, and sub-cloud will be made in this thesis. In the course of distinguishing forcing mechanisms, a distinction is made between in-cloud evaporation and sub-cloud evaporation. In-cloud refers to the region above the conventionally determined cloud base (the lifted condensation level in non-precipitating storms). Since cloud droplets evaporate extremely rapidly, they are almost completely absent in descending and warming air parcels (see reference list in Section B for cloud droplet measurements). Thus, a downdraft that is in the interior of a cloud will contain almost no cloud droplets. Any processes that are occurring in this downdraft (in-cloud evaporation of raindrops, for example), while geographically in-cloud, are not acting on a cloudy parcel.

There is some discussion relating to the location of downdrafts and precipitation shafts relative to visually observed clouds. Conventional wisdom (in low shear environments) pictures a rainshaft penetrating the visible cloud base at the bottom of a cell. There is speculation (Emanuel, pers. comm.) that rainshafts penetrate and pass outside of clouds at higher levels. Observations of tilted updrafts in tropical convection and precipitation shafts under high visible cloud bases in mid-latitude storms lend support to this theory. On the other hand, the author has observed virga penetrating directly from underneath conventional looking cloud bases. Convincing observational evidence supporting either the conventional or Emanuel viewpoints would be useful.

Downdrafts penetrating into the outside environment well above the conventional cloud base would experience the outside environment for a longer time than downdrafts penetrating the lowest regions of the cloud. In the former downdrafts, sub-cloud evaporation as described by Srivastava (1985) would be more significant. The quality of air entrained into these downdrafts would also be different. The model simulations presented in Chapter 3 include differing entrainment environments (in-cloud and out-of-cloud) and thereby cover the range of viewpoints.

Except in references to other work, the term microburst has been avoided in this thesis. All the downdrafts studied here produced outflows that were, or could have been, termed microbursts. While there are claims that microbursts can be distinguished from ordinary outflows, no physics peculiar to microbursts, as distinguished from that involved in ordinary outflows from ordinary thunderstorms, was involved in these four downdrafts.
CHAPTER 2 Analysis Techniques

A. Introduction

This Chapter discusses the techniques used in analyzing the observations of three downdrafts in Huntsville and the downdraft in Orlando. The Huntsville and Orlando radar arrays are compared to previous multiple Doppler radar arrays. Wind field variances are calculated and compared and various techniques of obtaining vertical velocities near the ground are discussed. The processes involved in converting sparse radial data into filled Cartesian grids and accounting for storm motion are discussed.

B. Radar Analyses

1. General Description of Radar Arrays and Data

The downdrafts studied in this work occurred in Huntsville, Alabama during the summer of 1986, and in Orlando, Florida, during the fall of 1990. There were five Doppler radars operating in Huntsville and three in Orlando. Figure 2.B.1a shows the radar array in Huntsville. This radar array was the most densely packed multiple Doppler network to date. The baselines between the radars ranged from 8 km to 25 km, with a mean of 12 km. The radar array in Orlando was not quite as closely spaced, as shown in Figure 2.B.1b, with baselines of 17, 19, and 20 km. The geometry of the arrays was of primary importance in determining the analysis techniques that could be used.

Doppler weather radars transmit microwave radiation into precipitation systems and measure the intensity and the frequency shift of the returned radiation. If the wavelength of the radiation is large compared to the diameter of the precipitation particles, then the power returned from each particle is proportional to the sixth power of its diameter (Doviak and Zrnic, 1984). The power received at a radar from an atmospheric volume containing precipitation is proportional to the radar reflectivity, Z, defined as

\[ Z = \frac{1}{\Delta V} \sum D_i^6 \]

where \( \Delta V \) is the radar pulse volume being sampled, \( D_i \) is an individual precipitation particle diameter, and \( Z \) has units of \( \text{mm}^6/\text{m}^3 \). Typical values of \( Z \) range over several decades,
therefore the radar reflectivity factor is usually expressed in a logarithmic form, dBZ where 
\[ \text{dBZ} = 10 \log_{10} Z. \]

The returned radiation from each precipitation particle is shifted in frequency relative to the 
transmitted radiation due to the velocity of the particle relative to the radar (the well known 
Doppler effect). The shift in frequency is related to the motion of the particle according to 
the formula \( V_r = \frac{\lambda f}{2} \), where \( \lambda \) is the wavelength of the transmitted radiation, \( f \) is the 
frequency shift of the returned radiation, and \( V_r \) is the velocity of the particle relative to the 
radar. Since there are many particles in a radar pulse volume, a mean \( V_r \) is estimated from 
the spectrum of frequency shifts that is returned. Since the returned power is proportional 
to the sixth power of particle diameter, this mean velocity is biased towards the motions of 
the largest particles. Turbulence, windshear, and the distribution of particle terminal 
velocities in radar pulse volumes contribute towards causing a spectrum of returned 
frequency shifts. The typical standard deviation of radial velocity estimates has been 
estimated at 0.5 m/s (Lhermitte and Atlas, 1961) and 0.53 - 0.57 m/s (Doviak and 
Zrnic, 1984).

\( V_r \) is the mean particle radial velocity, not the mean air parcel radial velocity. To convert 
computed precipitation particle velocities into air velocities, precipitation particle terminal 
velocity corrections are employed (as discussed in Section 2.d, below).

2. Triple Doppler and Multiple Doppler Analyses

a. Dual Doppler Analyses

Dual Doppler analyses are common and the techniques need not be discussed in detail here. 
(See Ray et al., 1980 and Doviak et al., 1976) for a discussion of various methods and 
errors.) If a storm is sampled by two or more Doppler radars, the computed radial 
velocities can be used to determine two perpendicular components of the windfield. Since 
dual Doppler analyses techniques can retrieve only two perpendicular components (usually 
the horizontal components) of the wind field, the third component is usually determined by 
integrating the mass continuity equation using boundary conditions at the ground and/or at 
upper levels of a storm.
b. Rationale and Equations for Triple Doppler Analyses

Errors associated with the integration of mass continuity arise due to inaccuracies in the specification of the boundary conditions and because of the accumulation of random errors during integration. In vigorous convection with strong outflows, the assumption of zero vertical air velocity, \( w_a = 0 \), at the measured level closest to the ground is suspect. For example, Fujita (1985) describes a microburst with a vertical velocity of 12 feet per second (3 m/s) at an altitude of 300 feet (100 m). This is well within the typical lowest scanning volume of most radars and thus errors on the order of several meters per second may be introduced by setting \( w_a(z=0) = 0 \). Similarly, the setting of \( w_a = 0 \) at the cloud top is suspect, particularly in vigorous convection. In addition, the actual cloud top is not sensed by radar and an upper boundary condition must be applied to the highest level exhibiting measurable horizontal velocities. This is probably in a region where \( w_a \neq 0 \).

Ray et al. (1980) discuss many issues relating to multiple Doppler errors and show that errors associated with the boundary condition are less severe with downward integration because of air compressibility.

The direct retrieval of the full, three-dimensional, wind field would eliminate the sources of error noted above. If three or more radars sample a region of a storm it is possible to explicitly retrieve the complete particle velocity field. Lhermitte (1968) and Armijo (1969) first presented this technique; other discussions and extensions of the technique can be found in Lhermitte and Gilet (1976), Bohne and Srivastava (1976), Ray et al. (1978), and Ray et al. (1980). The reflectivity weighted (proportional to \( D^6 \)) particle velocity can be retrieved at any point by solving the system of equations:

\[
VR_i = U_p \cos(a_i)\cos(e_i) + V_p \sin(a_i)\cos(e_i) + W_p \sin(e_i) \quad i = 1, n
\]

(1)

where \( VR_i \) are the radial velocities measured by the \( n \) radars, \( a_i \) and \( e_i \) are the azimuth and elevation angles of the \( n \) radars, and \( U_p, V_p, \) and \( W_p \) are the Cartesian components of the particle velocity field. If measurements are available from more than three radars, the system is overdetermined and can be solved by minimizing error. The matrix inversion and solution is straightforward, lengthy to write, and has been presented several times elsewhere, beginning with Lhermitte (1968).
c. Velocity Errors Using the Triple Doppler Method

The errors associated with the triple Doppler networks can be calculated by using extensions of the method described in Lhermitte (1968). Lhermitte (1968) shows that the variance in the estimates of vertical velocities grows roughly as $1/z^2$, where $z$ is altitude, and become infinite in the plane of the radars (assuming, as is almost always the case, that all the radars lie in a horizontal plane, Raymond and Blyth (1989) being an exception, as discussed below). Since vertical velocity errors are dependent on the elevation angles with which the radars sample the storm volumes, small arrays like that at Huntsville and Orlando can produce very accurate fields at all but the lowest levels in storms. Widely spaced arrays can retrieve vertical velocities only at high altitudes.

Expected variance fields for the Huntsville and Orlando networks have been calculated. Normalized particle velocity variances for the Huntsville network at various altitudes are shown in Figure 2.B.2. The errors are normalized to a mean Doppler velocity variance of 1 m$^2$/s$^2$. The vertical velocity errors drop quickly with increasing altitude and are smallest near the radars, as expected. At 2 km altitude the contiguous region with less than 6 normalized variance units is over 100 km$^2$ in area and there are small regions with less than 2 units above each radar. Above 4 km altitude, the region exhibiting less than 2 units extends around all the radars for about 200 km$^2$. Since typical variances in radial velocity are about 0.5 m/s (Lhermitte and Atlas, 1961, Doviak and Zrnic, 1984), this array should theoretically allow the direct calculation of accurate particle velocities down below 2 km altitude in regions inside the area defined by the radars. The error fields for horizontal velocities ($u+v$) using overdetermined dual Doppler calculations ($w_p$ is assumed to be zero in the equation presented in the previous Section) are shown at the same altitudes. The errors grow slowly with height and are much smaller than the errors for vertical velocities up to very high altitudes. Figure 2.B.3 shows the individual u and v variance fields in the same region. The structure of these fields is complex but indicates that both components can be determined with great accuracy within the southern portion of the five radar array. In the north, near the CP2 radar (at coordinate $(x,y) = (94,148)$), the $u$ component is poorly resolved as would be expected with this particular geometry. Figure 2.B.4 is a vertical cross Section through the $w_p$ variance field showing the vertical structure. The errors grow rapidly near the ground except directly over a radar located at the right edge of the plotted slice. The figure indicates the small errors expected for altitudes above 2 km. Figures 2.B.5 shows the vertical velocity variances for the Orlando network at the same
altitudes as shown for Huntsville in Figure 2.B.2. At 2 km, the region of 6 units or less variance is confined to small regions above each radar. Only at altitudes above 3-4 km does a contiguous region of low expected variance exist.

Figure 2.B.6 shows the relative geometry of several arrays that have been used for multiple Doppler studies. Previous studies, including Ray et al. (1978), Miller (1978), Harris and Fankhauser (1978), Foote and Frank (1983), and Caracena and Maier (1987), have been conducted with arrays that have been considerably larger than the arrays used in Huntsville and Orlando. The use of long baseline arrays led Ray et al. (1980) to conclude that direct retrieval was inferior to various integration techniques. Lhermitte and Williams (1985) and Comes and Krehbiel (1986), using dense but not overdetermined networks, found that direct triple Doppler derived precipitation velocities could be obtained. Raymond and Blyth (1989), using a fairly dense, and overdetermined network, were able to calculate precipitation velocities directly. This latter network was unique in that one of the radars was elevated by 1200 m above the plane of the other three. Because of this the vertical velocity variances never became infinite. But, since the elevated radar is only 1200 m above a network that is approximately 30,000 m across, the variances are not significantly different than those from a hypothetical similarly dimensioned coplanar array. Vertical velocity variance fields for several arrays are shown in Figures 2.B.7 and 2.B.8. In the former figure, the fields are shown for a horizontal layer at 2.4 km altitude. A large fairly convex region of low variance (< 4 units) exists only in the Huntsville array. Both the Huntsville and Orlando arrays, and to a lesser extent the Harris and Fankhauser (1978) array, contain convex regions with variances of less than 10 units. In all the other arrays, regions with acceptably low errors exist only immediately above, or in thin lines between, the radars. At 9.6 km altitude, as shown in Figure 2.B.8, all the arrays can retrieve reasonably accurate vertical velocities, although the Huntsville and Orlando arrays still provide the most favorable expected results.

The details of loading the radially acquired radar data into Cartesian grids and the shifting of the fields to account for storm motion are discussed below in Section C. Since the data have been shifted in the Cartesian grids to account for storm motion, the apparent locations of the radars used to calculate the trigonometry terms in the above figures have also been altered so that the original geometric relationships between data locations and radars have been preserved. This causes distortions in the array that alter the variance fields presented above. In addition, the scanning strategies used in Huntsville were uncoordinated in both space and time; many locations were sampled by only three or four radars out of the
possible five. This results in variance fields with higher values than those presented above. This was not a problem in Orlando, where coordinated scanning was conducted.

Figure 2.B.9 shows variance fields at 2 km altitude for a particular volume scan made from radar data taken near 2031 Z 11 July 1986 in Huntsville. Panel a shows the actual vertical velocity variance field. Panel b shows the theoretical vertical velocity variance field assuming that five radars scanned everywhere. Panel c shows the number of radars that actually scanned in different locations. In the region extending from the center of the plot area to the northwest (top left), five radars scanned and the actual variances are very close to the five scanning radar values. The minor differences are due to storm motion corrections which compute the computational radar array. In the regions where only four radars scanned, the variances are two to five times higher than those predicted for five radars. The regions where only three radars scanned have extremely high variances. (Extremely large variance values cause overflow conditions in the analysis software, resulting in the noisy plotted pattern.) These regions are outside of the triangular area defined by the three scanning radars so direct vertical velocity retrieval is almost impossible.

d. Determinations of Precipitation Terminal Velocity

To calculate air parcel vertical velocities from the measured particle velocities, the $w_p$ estimates determined with equation (1) must be corrected for precipitation terminal velocities. It is assumed that all precipitation particles are moving with the surrounding air parcels except for a correction due to the particles' terminal velocity. As particles pass through air parcels with differing velocities, they adjust to the above state very rapidly, within about 20 m (Pruppacher and Klett, 1980). Since precipitation terminal velocities, $w_t$, are not directly observed, radar reflectivity-terminal velocity relations are usually employed. In this work, the terminal velocity formulation of Pasqualucci (1975) is used. His empirical formula, derived from vertical measurements in rain provides results very similar to others (Joss and Waldvogel, 1970, Rogers, 1964). Comparisons of these formulations are shown in Figure 2.B.10. The plotted relations are:

\[
\begin{align*}
  w_t &= 2.67 Z^{0.0095} \\
  w_t &= 2.6 Z^{0.0107} \\
  w_t &= 3.8 Z^{0.0071}
\end{align*}
\]

Pasqualucci (1975)  
Joss and Waldvogel (1970)  
Rogers (1964)
They all predict terminal velocities between 8 m/s and 9 m/s (at sea level) at 50 dBZ reflectivity. The values drop quickly at low reflectivities. The formulas are suspect near the cloud edge, since they are incapable of producing results for reflectivities of less than 0 dBZ because they are exponential relations in Z. Other formulations, such as Dyer (1975), \( w_t = 5.18 \times 10^{-9.03} \), shown in Figure 2.B.10, have been derived for hail and snow, but, as will be seen, the downdrafts studied here contained predominantly liquid precipitation. Had the downdrafts occurred in ice phase regions, the employed terminal velocity calculations would have to have been modified to account for graupel, snow and hail and the confidence inspired by the consistent rain results discussed above would be reduced.

In this work, the Pasqualucci (1975) terminal velocities have been adjusted for the decrease of air density with height, \( z \), by using the relation

\[
wt (z) = wt (z=0) \left( \frac{\rho (z=0)}{\rho (z)} \right)^{0.5}
\]

where the air density \( \rho (z) \) has been calculated by assuming a ground temperature of 300 K, a lapse rate of 7 K/km and a pressure scale height of 8 km.

Tests of the assumed precipitation particle terminal velocity formulas can be conducted by combining dual and triple Doppler techniques. The difference between the dual Doppler vertical air velocity, \( w_a \), results and the triple Doppler vertical particle velocity, \( w_p \), results provides an estimate of precipitation particle terminal velocities, \( w_t \). This result is independent of any assumptions concerning precipitation phase. Figure 2.B.10b shows the terminal velocities predicted by the Pasqualucci (1975) formulation and the method just described, as well as the difference between the two results. The fields have been filtered to eliminate data from regions that exhibit radar reflectivities below 20 dBZ and to eliminate data from regions in which the dual Doppler and triple Doppler fields disagree by more than \( \sqrt{5} \) m/s. (In those regions either the triple Doppler \( w_p \) field or the dual Doppler \( w_a \) fields are probably in error.) The difference in the estimates of \( w_t \) are quite small, generally less than 1 m/s through a broad region, suggesting that the Pasqualucci (1975) formulation and the vertical wind calculations are quite accurate.
Three dimensional Divergence Used to Assess Triple Doppler Reliability

Since direct triple Doppler calculations derive the complete, three-dimensional, wind field, it is possible to calculate the full, three-dimensional divergence, \( \nabla \cdot \mathbf{V} \).

Following Holton (1979), mass continuity provides that:

\[
\frac{1}{\rho_0} \left( \frac{\partial \rho'}{\partial t} + \mathbf{V} \cdot \nabla \rho' \right) + \frac{w_a}{\rho_0} \frac{\partial \rho_0}{\partial z} + \nabla \cdot \mathbf{V} = 0
\]


\begin{align*}
\text{A} & \quad \text{B} & \quad \text{C} \\
\frac{1}{\rho_0} & \frac{\rho'}{\partial t} & + \mathbf{V} \cdot \nabla \rho' & + \frac{w_a}{\rho_0} & \frac{\partial \rho_0}{\partial z} & + \nabla \cdot \mathbf{V} = 0 \\
\end{align*}

where \( \rho_0 \) is mean air density, \( \rho' \) is the perturbation about that mean, and \( \mathbf{V} \) is air velocity.

Scale analysis of the above equation can be conducted by choosing generous values for the variables that may exist in the storms studied here: \( \rho'/\rho_0 = 10^{-2} \) consistent with 3 K temperature perturbations in a 300 K environment, \( V = w_a = 10 \text{ m/s} \) consistent with the observed velocities in the storms studied here, \( \frac{\partial \rho_0}{\partial z} = 1/8000 \text{ m} \) consistent with an 8000 m density scale height, \( t = 200 \text{ s} \) a short time for significant changes in the density field and \( L = 500 \text{ m} \), a small distance scale over which these density changes might occur.

Term A becomes, when divided by \( \rho_0 \), \( 5 \times 10^{-5} \) \text{/s}, term B becomes, when divided by \( \rho_0 \), \( 2 \times 10^{-4} \) \text{/s}, term C becomes \( 10^{-3} \) \text{/s}.

Thus, even assuming the rather extreme values in the above scaling, \( \nabla \cdot \mathbf{V} \) should be less than or equal to \( 10^{-3} \) \text{/s} if the three-dimensional winds are calculated without error. As discussed above, errors can arise from random errors associated with individual Doppler radial wind estimates, or from other sources including errors in correcting for storm motion, storm evolution, Cartesian gridding and interpolation, and terminal velocity estimation. Comparisons of theoretically predicted \( \nabla \cdot \mathbf{V} \) and measured \( \nabla \cdot \mathbf{V} \) give an indication of whether the errors in the analyses are dominated by expected random Doppler error or by other sources, and also indicate whether the triple Doppler winds are reliable.

Figure 2.B.13 shows a vertical slice through a region including cell F at 2031 Z. Panel b displays smoothed observed \((\nabla \cdot \mathbf{V})^2\) (the divergence has been squared in order to produce positive definite values). Panel c displays the expected vertical velocity variance field, \( \sigma_w^2 \). Panel a shows the predicted \((\nabla \cdot \mathbf{V})^2\) using the relation.
\[(\nabla \cdot \mathbf{V}) \text{ (m/s/gridpoint)} = \sqrt{2\sigma_w^2} / \text{(2 gridpoints)} \]

The most noticeable feature of the \((\nabla \cdot \mathbf{V})^2\) fields is that they are small aloft and very large near the ground. At any given altitude, the values are smaller in the region sampled by five radars than in the region sampled by only four. The first shading contour interval is at \((.005/s)^2 = 2.5 \times 10^{-5} /s^2\), which is much larger than the square of the value predicted from the scale analysis of the mass continuity equation presented above, since \((10^{-3}/s)^2 = 10^{-6} /s^2 \ll 2.5 \times 10^{-5} /s^2\). Therefore the plotted values are measurements of analysis error, not atmospheric deviations from \(\nabla \cdot \mathbf{V} = 0\).

The agreement between the predicted and measured \((\nabla \cdot \mathbf{V})^2\) is very close through most of the slice. (Near the ground, the analysis software imposed an artificial upper limit of 4.0 on the calculated values. This is in the region of very high error and differences between values that are over 4.0 are not important.) The observed errors do not exhibit a simple monotonic z dependence at small scales however. This is not surprising since the actual errors include uncertainties in terminal velocity, motion errors, evolution errors and gridding and interpolation errors. In a slice through cell R, Figure 2.B.14, there is large \((\nabla \cdot \mathbf{V})^2\) near the melting level in the highest reflectivity, near (85,20), possibly due to bad terminal velocity estimation in the presence of enhanced reflectivity caused by melting ice phase precipitation. Large \((\nabla \cdot \mathbf{V})^2\) values in the high reflectivity gradient regions are probably due to terminal velocity errors and storm evolution errors. There are also large values at high altitudes, above 7 km, in regions where ice is the predominant phase and the Pasqualucci (1975) terminal velocity relation is suspect. In this slice, the region where four radars scanned does not exhibit reduced \((\nabla \cdot \mathbf{V})^2\), compared to the thrice scanned area, because the fourth radar was poorly located relative to the plot area in terms of triple Doppler geometry.

The observed values of \(\nabla \cdot \mathbf{V}\) provide an upper bound on possible violations of mass continuity in the atmosphere. While not surprising, the small observed values of \(\nabla \cdot \mathbf{V}\) do provide comforting confirmation that, on short time scales, the mass in the atmosphere is, in fact, conserved (at the limit of detectability). This is assumed in most Doppler radar studies as well as in many other endeavors in meteorology, but has seldom, if ever, been directly observed on scales larger than the laboratory.
Mixed Triple and Dual Doppler Analysis Techniques

Since the errors in directly retrieved vertical velocity fields are very large near the ground, integration of the mass continuity equation must be used to obtain accurate results. Two types of integration have been used, upwards and downwards, described below. In all integrations, terms in which the vertical gradient of density appears are included. Terms including only horizontal and temporal variations of density are neglected. Thus, the vertical velocity at any level in the grid \( zn+1 \), is calculated from the value at level \( zn \) using the following equation:

\[
w_a(z_{n+1}) = w_a(z_n) \left( 1 \pm \frac{\Delta z}{H} \right) \pm \left( \frac{du_{n+1/2}}{dx} + \frac{dv_{n+1/2}}{dy} \right) \left( 1 \pm \frac{\Delta z}{2H} \right) \Delta z
\]

where \( H \) is the density scale height of the atmosphere, \( \Delta z \) is the vertical grid spacing and the signs of the terms are determined by the direction of the integration.

i. Upward integration from the ground

The first approach is the use of the traditional upward integration from the ground assuming a boundary condition of \( w_a(\text{ground}) = 0 \). The integration errors associated with this type of integration are shown for the two networks in the vertical slice of Figure 2.B.11. The errors grow rapidly with height and are soon larger than the triple Doppler errors discussed above. This comparison is complicated, however, by the existence of initial and boundary condition errors in the latter fields and by errors in precipitation particle terminal velocity estimates in the former.

ii. Application of a Mid-Level boundary Condition and Downwards Integration

The other integration technique utilizes a mid-level boundary condition in a region where \( w_a \) is well determined; vertical winds are determined by the integration of mass continuity downwards to the ground. While previous workers, Ray et al. (1980) for example, have used downward integration from the uppermost regions of storms, this is believed to be the first use of a mid- or low-level boundary condition in this type of analyses. It is only feasible here because of the unprecedentedly accurate values of triple Doppler vertical air velocity that can be calculated using the Huntsville (and to a lesser extent, the Orlando) data. The technique produces the best results at mid- to low-levels in storms. Ideally, the
computation produces \( w_a(\text{ground}) = 0 \); deviations from zero indicate the level of accumulated integration and boundary condition errors, as discussed in Appendix 4.

The errors associated with a typical downward integration of mass continuity using a mid-level boundary condition in the Huntsville array are shown in the vertical slice of Figure 2.B.11. In the displayed fields, the boundary level was chosen to be the surface on which \( \sigma^2_w = 4 \text{ m}^2/\text{s}^2 \). The errors increase rapidly as the integration proceeds downward, but the expected errors are less than that in the directly calculated fields shown in Figure 2.B.2. In Huntsville, where the radar array is very dense and overdetermined, the switchover altitude could be as low as 2 km. In other arrays it must occur at a much higher level. In the large radar arrays discussed earlier, the switchover altitude is so high that a \( w = 0 \) boundary applied at cloud top is used. Raymond and Blyth (1989), with a somewhat smaller and overdetermined array, found that they could use the directly calculated \( w_a(\text{cloud top}) \) as a boundary condition. They used the directly calculated precipitation velocities through the interior of storms for calculating precipitation trajectories.

Techniques combining both upwardly and downwardly calculated velocities at low levels were found to produce only marginal gains in accuracy since, empirically, the downward integration from mid-levels was quite accurate and produced small \( w_a(\text{ground}) \) residuals (see Appendix 4). The choice of a constant altitude boundary level or a constant \( \sigma_w \) boundary level also had no significant impact on the analyses because \( \sigma_w \) surfaces were fairly level over the small regions in the interior of the radar array covered by any particular outflow.

The choice of the triple Doppler-integrated dual Doppler switchover level (between direct and integrated calculations) was important. Two methods were used to determine this level and were also used in the Orlando data to filter regions with bad data. The first method was to use the theoretically calculated variances to determine a surface below which the vertical winds would be unreliable (\( \sigma^2_w = 4 \text{ m}^2/\text{s}^2 \) as above, for example). The second, and preferred, method was to use the measured values of \( \nabla \cdot V \) to determine this level, as discussed in the previous Section.
g. Differential Reflectivity Measurements

In order to diagnose the possible role of the melting of ice as a forcing mechanism in the
downdrafts, the presence and amount of ice in the cells must be measured. This was done
with the aid of differential reflectivity measurements, ZDR, taken by the CP2 radar. As
discussed in Seliga and Bringi (1976) and Doviak and Zrnic (1984), liquid raindrops are
shaped like oblate spheroids. The principal axes of the oblate spheroids are nearly vertical.
Since the radar backscattering cross-section from any raindrop (assuming Rayleigh
scattering) is proportional to the sixth power of drop radius (as measured along an axis
parallel to the radar polarization), small differences between the length of the horizontal and
vertical axes can be detected from differences in the returned power from horizontally and
vertically polarized incident radiation. Large raindrops, with diameters of 2 mm, typically
exhibit ZDR values of about 2-3 dB.

Solid phase precipitation is often more anisotropic than liquid raindrops (conical graupel,
ice columns, snow dendrites) and thus one might expect very large ZDR signatures from
ice. In fact, many individual ice particles can reflect with a large ZDR. The ice phase
precipitation typical in thunderstorms, graupel, often tends towards rough sphericity,
however. More importantly, it tends to tumble and thus exhibits rapidly varying ZDR
values. When integrated over a radar pulse volume containing a great number of randomly
tumbling particles, the net ZDR is very nearly zero. An exception to this can occur in
quiescent regions where ice needles or plates are common. Since these particles can exhibit
preferred orientations, net ZDR can result. Melting snow can also exhibit large values of
ZDR.

In summary, regions that exhibit large values of ZDR, above about 1-2 dB, contain
predominantly large raindrops while regions with almost zero ZDR are dominated by ice
phase particles or small, less than 1 mm diameter, raindrops.

h. Reflectivity-Mass Relationships

In order to diagnose the magnitude of precipitation loading, evaporation, and melting that
are occurring in the storms, it is necessary to know the amount of precipitation mass that is
available to these processes. There were no direct measurements of the mass inside the
cells discussed in Chapters 3 and 4. Radar reflectivity measurements can be used in lieu of
these by employing empirically developed relationships, commonly termed Z-M
relationships. The formula proposed by Geotis (1971) is used in this study. It was derived from measurements of drop size distributions and is consistent with a relationship that can be computed from the Marshall and Palmer (1948) drop size distribution formula. The Geotis (1971) relation and several others are plotted in Figure 2.B.12. The plotted formulas are:

\[
\begin{align*}
Z &= 21000 \ M^{1.43} & & \text{Geotis (1971)} \\
Z &= 20410 \ M^{1.47} & & \text{derived using Z-R and M-R relations derived from the dropsize distribution in Marshall and Palmer (1948)} \\
Z &= 23900 \ M^{1.82} & & \text{Douglas (1964) for rain} \\
Z &= 26400 \ M^{1.82} & & \text{derived using revised Z-R and M-R relations from Marshall et al. (1947) from Marshall and Palmer (1948)} \\
Z &= 26036 \ M^{1.64} & & \text{derived using Z-R from Wexler (1947) by Marshall and Palmer (1948)}
\end{align*}
\]

These relations, which all apply to rain, are in qualitative agreement. Below 45 dBZ and 1 g/m³, they agree to within about 15%. At the highest reflectivities, the relations of Geotis (1971) and Marshall and Palmer (1948), which are in extremely close agreement, grow the fastest so that at 55 dBZ, they predict 7 g/m³ while the other relations predict only about 3 g/m³. The Geotis (1971) relation was well suited to the modeling effort of this study since it is consistent with the Marshall and Palmer (1948) dropsize distribution from which the evaporation scheme, detailed in Appendix 2, was derived. Following Geotis (1971) and Geotis (pers. comm.) the relation is not extended to regions in which the reflectivity exceeds 55 dBZ.

These Z-M relations are valid only in rain. The different reflectivity characteristics and particle size distributions of snow, graupel, and hail can cause these relations to break down. Douglas (1964) provides several relations that apply in hail and snow and Battan (1959) presents several different relations that apply in a range of varied weather situations. These Z-M relations determine only the precipitation water content and not the total water content. The cloud droplet water content, which does not contribute to Z, is not included. Typical cloud droplet water contents are near 1 g/m³ (see references in Chapter 1, Section B).
C. Gridding, Interpolation and Shifting of Radially Acquired Doppler Data

Radar data are necessarily recorded in a polar format. Also, a finite time is required for a radar to sample a volume of space. Because of these factors, the raw data must be manipulated in order to reduce it to Cartesian grids that represent approximate snapshots in the history of a storm.

Gridding of the data for this study was conducted in a straightforward fashion. The details of the gridding process did not significantly affect the derived fields and physics. The dimension of the gridboxes were always 200 m on a side, corresponding roughly to radar pulse volumes and much smaller than the physical features that are diagnosed. Radar gates were loaded into most all gridboxes into which they intruded. This was accomplished by determining the location of fifteen points along the edges and central line of each radar pulse volume. These were arranged into three "X" patterns at the inner surface, center, and outer surface of the volumes as shown in Figure 2.C.1. Multiple hits to particular gridboxes could occur from the fifteen points in a single gate and from points in adjoining gates. Weighted averages were calculated when these multiple hits occurred. Filtering was sometimes applied so that data taken at times closest to the nominal time of the volume scan would be preferentially treated either with exponential weighting or thresholding.

Since storms tended to move at significant velocities during a volume scan (The Huntsville storms that are discussed in Chapter 3 moved at about 10 m/s. The Orlando storm discussed in Chapter 4 moved at 3 m/s.), it was necessary to shift the data to account for storm motion. The storm velocity was subjectively determined by examining several reflectivity fields that were gridded with different motion corrections. Difference fields were created by subtracting reflectivity fields made from volume scans at adjacent times. The assumed storm velocity was varied about the first guess field and the resultant difference fields were compared. The final storm velocity was chosen by subjectively minimizing the largest scale differences. The residual differences were due to storm evolution, for which corrections were not made.

Since most radars scan too slowly to sample a complete volume or even an appreciable section of a complete volume during the course of a few minutes, they usually employ a sparse scanning strategy. The result can be large gaps in azimuth and elevation between tilts. The gaps from each individual radar's scanning strategy are located in different locations and thus the intersection of grid locations that are actually sampled by three or
more radars is a collection of isolated points. Since interpretation and analyses of such a set of points would be difficult and probably fruitless, grid filling must be conducted. As with the grid loading technique, the specific method of grid filling should not significantly influence the fields or inferred physics. The method used here was inversely weighted trilinear interpolation with a threshold searching distance. When the filling algorithm found a grid location with no data, it searched in the $x$, $y$ and $z$ directions, both higher and lower, until a loaded value was found or until a threshold distance was exceeded. The empty value was set to the weighted average of the linear interpolation between the loaded values in the three directions. The weighting factors were equal to the inverse of the total distance between the loaded values in each direction.

Figure 2.C.2 shows a horizontal slice through a typical gridded volume scan comprised from a series of RHI's* with an inter-tilt spacing of 5 degrees. The pre-filling inter-tilt gaps are about 600 m in the plotted region. The time difference between the start and end of the scan is about 2.5 minutes. The post-filled field appears, to the eye, to be absent of any banding artifacts arising from the filling process.

When the gridded fields from up to five radars are combined to calculate triple Doppler and overdetermined dual Doppler wind fields, the amount of storm motion shifting that occurred during loading is subtracted from the radar locations used when applying the triple Doppler equations. Thus, the true relationship between the locations at which radial velocities were measured and the radars that measured them are preserved.

Storm evolution limits the accuracy of gridded volume scan data. The accuracy is difficult to calculate and there is no simple expression for the time at which a particular triple Doppler wind calculation might apply. Since, particularly in Huntsville, each radar had a different scanning strategy, the times at which data were taken at particular grid locations varied in an almost capricious manner. Figure 2.C.3 shows the reflectivity and time difference fields from the five radars for data taken during a three to four minute period. The time difference fields show, that at many locations, observations by the different radars were separated by up to three minutes. For example, the location (60,68) was observed at the following times, CP2: 2029.4 Z, CP3: 2032.4 Z, CP4: 2030.8 Z, FL2: 2033.7 Z, and UND: 2033.3 Z. Hence, the UND measurement came 3.9 minutes after CP2 scanned the

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* RHI from range-height indicator designates a radar tilt in which the azimuth angle is held constant while the elevation angle is varied, resulting in a vertical slice through the storm.
region. Scan strategies also varied significantly from radar to radar (and sometimes from scan to scan at a particular radar). CP2 and CP3 took RHI's. CP4 took PPI's* with an inter-tilt spacing that was so small that they very nearly filled the volume with measured data. (No filling was necessary.) FL2 took PPI's but in an interleaved fashion that caused adjacent grid locations to contain data from very different times. UND recorded two PPI tilts and then several RHI's.

Comparisons between the gridded reflectivity fields from the five radars show similarities and some differences. The measured large scale features of the reflectivity field are similar, although the CP3 measured values are lower than the rest. The reflectivity protrusion in the northwestern portion of the cells is evident in all the radar grids except UND’s, where it is only suggested at the edge of the scan region. Very small storm features are washed out, however. The reflectivity notch in the southwestern portion of the storm is clear in the CP3, CP4, and FL2 grids, but is absent in the CP2 grid, and only hinted at in the UND grid. CP2 never took measurements in the notch region and the linear interpolation algorithm averaged the higher values that were nearby across the notch. Any velocity perturbations that were peculiar to the notch region have thus been lost. This analysis is somewhat complicated by the fact that the CP2 observations in the notch region were taken about three minutes earlier than the observations by the other radars. Perhaps the notch evolved during the period between the CP2 observation at 2029.7 Z and the next observation, at 2031.7 Z, by CP3.

* PPI from plan-position indicator designates a radar tilt in which the elevation angle is held constant while the azimuth angle is varied, resulting in a conically shaped scan pattern.
CHAPTER 3  Huntsville Downdrafts

A. Introduction

A cluster of thunderstorms passed through the radar array in Huntsville, Alabama, on 11 July 1986. Three of the storms produced downdrafts and outflows inside the array. In this Section, these downdrafts are described and the forcing mechanisms that caused them are diagnosed. The downdrafts are described using multiple Doppler analyses aloft and at the surface. Surface mesonet thermodynamic observations are used to aid in distinguishing deep from shallow outflows. Simple model calculations are used to diagnose individual forcing mechanisms. These calculations reproduce the observed radar and surface characteristics of the downdrafts and outflows.

B. Synoptic Setting

The large scale environment in which the Huntsville downdrafts developed was typical for mid summer in the southeastern United States. The mid-level and surface flows were from the southwest as shown in the 500 mb and surface dataplots valid at 1200 Z, about eight hours before the downdrafts occurred, Figure 3.B.1. Temperatures, dew points and \( \theta_e \) values were unremarkable except for the bogus dew point observations that were common at 500 mb. (Dew point depressions of over 30 K indicate relative humidities below the instrumental threshold sensitivity of 20%.) A balloon sounding was taken at the Redstone Arsenal, just southeast of the study area, at 1800 Z, about two hours before the outflow events. This sounding, Figure 3.B.2, is also unremarkable. The lifted condensation level of undilute surface air was at about 900 mb, and the level of free convection was 880 mb. There was potential for convection as indicated by the 500 mb surface lifted index of -5 K. The winds were light (\(< 3 \text{ m/s}\) ) from 240° at the surface and freshened to 15 m/s at 750 mb from the same direction. At 500 mb, the winds weakened to 12 m/s from 250°. The profile of \( \theta_e \) is shown in Figure 1.3. It is presented as a typical summertime sounding and exhibits a broad minimum of \( \theta_e \) near 600 mb.

Convection was widespread across the southeastern United States during the afternoon as the National Radar Summary valid at 2035 Z shows in Figure 3.B.3. The cluster of storms that produced the studied outflows is not distinctly discernible but is in the region of convection to the north of the most intense radar echoes visible in central Alabama. The National Radar Summary indicates that the radar cloud tops were at 45000 feet (13.7 km)
in cells to the south of the study area; the local radars indicate that radar echo tops in cells within the network are between 13.5 km and 14 km.

C. Storm History

Soon after 2000 Z, a cluster of thunderstorms moved into the study area from the west. At least four separate outflows were produced by this cluster. Figure 3.C.1 shows the surface divergence field at 2031 Z, when all three studied outflows are occurring. The cells associated with each outflow are labelled on Figure 3.C.1 as cell F (front), cell R (rear), and cell S (south). A detailed description of each downdraft and its forcing mechanisms follows in Section D.
D. Detailed discussion of individual downdrafts in Huntsville.

1. Cell R

a. Introduction

This Section describes the structure and physics of the downdraft in the cell R. The location of the cell R outflow relative to the other cells is shown in Figure 3.C.1. The downdraft had origins at mid-levels, near 4-5 km altitude. It was forced primarily by precipitation loading and melting of raindrops, as the observations and model results will show.

b. Cell R History

The time history of the vertical development of reflectivity in this cell is shown in Figure R.1. The cell develops rapidly before 2016 Z. Reflectivities above 50 dBZ reach 6 km for a brief time from 2016 Z to 2019 Z. After that time, this high reflectivity core drops to 3.8 km at 2031 Z. Reflectivities of 40 dBZ rise to an altitude of 9 km by 2022 Z and then collapse rapidly to 5.5 km by 2031 Z. There are very high values of reflectivity, near 60 dBZ, near the melting layer, at 2022 Z. These values are probably due to melting precipitation. This reflectivity history suggests that moderate precipitation water contents were present at mid-levels in the cell: 3 g/m³ up to 6 km. These estimates are based on a Z-M relationship presented in Geotis (1971) discussed in Chapter 2. The estimates below 6 km are reliable because in much of the cloud below this level the predominant water phase was liquid as will be shown later when ZDR measurements are presented. Above 6 km, the Z-M relation is not reliable. Approximately 3 g/m³ ice phase precipitation content up to about 9 km can be inferred by noting the precipitation water content that is present just below the melting layer. The impact of this mass level on the dynamics of the downdraft will be discussed in Section f.

Radar CAPPIs* depict the evolution of the outflow producing cell during the analysis period. Cell R develops quickly at the rear of cell F as can be seen in Figures R.2a and R.2b. (See Figure 3.C.1 for the geographical relation between cell R and cell F.)

* A CAPPI is a constant altitude plan-position indicator plot. Here, they are horizontal slices through the gridded radar data.
The figures show horizontal slices through the reflectivity field of the cell, at 600 m and 4 km altitude, at various times during its evolution. The time history of the low altitude reflectivity in cell R is more difficult to determine, particularly at the earliest times, due to the proximity of cell F, just to the east (right) of cell R. It becomes identifiable only at 2022 Z in Figure R.2a, near (54,76). By 2025 Z, it is clearly separate from cell F and exhibits reflectivities of 50 dBZ by 2028 Z, when it is located at (50,83). By 2034 Z it has shrunk considerably. Aloft, at 4 km in Figure R.2b, the cell is identifiable as an elongation of reflectivity at 2016 Z. It is more clearly separate at 2019 Z and is stronger than cell F at 2022 Z when peak reflectivities are near 55 dBZ, at (50,76). The cell fades rapidly and is difficult to identify by 2031 Z.

The surface divergence grows rapidly as the cell and downdraft develop. Figure R.3 shows surface divergence along with contours of reflectivity from the 800 m level (since the surface level reflectivities are contaminated by ground clutter and beam cutting effects). The divergence grows rapidly from 2025 Z, peaking at 2028 Z and then decaying somewhat by 2031 Z. Peak values at 2028 Z are .011 /s. (Unsmoothed values peak at .015 /s.) The standard measure of outflow strength, the differential radial velocity, \( \Delta V \), peaks around 2025 Z, as illustrated in Figure R.4.

The ground relative winds in the outflow are almost all westerly, but in the following discussions, the cell motion will be subtracted to obtain cell-relative windflow. The cell motion was \((u,v)=(10,3) \) m/s. Figure R.5 shows vector fields of the cell-relative and ground-relative wind fields. The region of strong divergence near coordinate (50,87) is evident.

c. Surface Mesonet Observations Below Cell R

Surface thermodynamic measurements of the downdraft outflow are used to diagnose the forcing mechanisms and depth of the downdrafts. Figure R.6 shows a portion of the Huntsville surface mesonet array, with the position of the the four southernmost radars shown. (CP2 was several kilometers north of the mesonet array.) The labelled gridpoint locations correspond to the radar grid locations in all figures, normalized to the shifted location at 2028 Z. Radar grid locations for other times are written above the plot. (Since the radar grid is shifted eastward with the storms, the ground and therefore the mesonet array shifts to the west with time.) The locations of the surface divergence at 2025 Z, 2028 Z, and 2031 Z, corresponding to the times shown in Figure R.3, are indicated. The
radar-measured outflow crosses station 104 at approximately 2029 Z - 2032 Z, and grazes station 24 (P24) at 2031 Z. The records from these two stations are shown in Figures R.7 and R.8.

Station 104 is affected by both cell F and cell R; the record shown in Figure R.7 shows the effects of both. The dramatic changes in the record that occur during the period from 2020 Z until 2028 Z are discussed along with cell F in Section 2. The period of interest here is from 2028 Z until 2034 Z, during the passage of cell R. At the beginning of this period, the temperature is depressed and the relative humidity and vapor density are high due to the effects of the outflow from cell F. The value of $\theta_e$ is high, at 350 K, very nearly the value before the impact of cell F. At 2028 Z, the average wind speed has dropped to a relatively quiescent 4 m/s. During the passage of cell R, the wind speed increases to 9 m/s, the temperature drops from 298 K to 296 K, and notably, $\theta_e$ drops from 352 K to 340 K and the water vapor density drops from 21 g/m$^3$ to 18.5 g/m$^3$. This desiccation, associated with a slight relative humidity drop, occurs during a period of heavy, 60 mm/hr, rainfall. The measured mesonet values have been adjusted slightly to account for the presumed wetting of the relative humidity instrument during the passage of cell F. This adjustment is discussed more fully in Appendix 3.

Station 24 is affected by cell R from about 2031 Z until about 2034 Z, as shown in Figure R.8. It is more difficult to separate the effects of cell F from cell R at this location partially because the outflow from cell R had weakened by this time. At about 2026 Z, cell F causes the temperature to drop and relative humidity and water vapor density to rise. When the remnants of the cell R outflow pass there is no windspeed increase; the rain associated with the earlier cell tapers off. The value of $\theta_e$ and vapor density drop in a manner similar to station 104, although the magnitude of the drop is less. These drops also occur during significant, 60 mm/hr, rainfall.

It is this surface behavior, particularly the desiccation and $\theta_e$ drops during heavy rain, that will have to be explained by examining the radar evidence aloft, in Section d, and the constrained model results, in Section f.

d. History of Cell R as Deduced from Triple Doppler Analyses

Radar data were loaded into three-dimensional Cartesian grids representing snapshots of three minute duration. Thus, there were grids corresponding to the nominal times of
2022 Z, 2025 Z, 2028 Z, 2031 Z, and 2034 Z. This period was roughly the update time of the radars, with the exception of UND, which scanned more irregularly.

i. 2022 Z

As the cell moved into the radar array, triple Doppler analyses became possible, particularly after 2022 Z. Radar analyses at 2022 Z show that a deep downdraft exists, extending to at least 3.2 km, and that engulfment of air is occurring at 3-4 km above the ground.

The upper region of the downdraft can be seen in Figure R.9. This figure displays the same horizontal region as the reflectivity CAPPI's in Figure R.2a. The slice extends for 5 km x 5 km at 3.2 km altitude. In panel a, the horizontal convergence field, shaded, is shown along with vectors of horizontal wind. Reflectivity is contoured. Engulfment of environmental air is occurring in a broad region at the southwest of the cell (as confirmed by tracing air parcel streamlines). There is an area of strong convergence centered at roughly (50,75) in the engulfment region. The cell was still outside the radar array and therefore vertical velocity calculations are suspect. There is evidence of the downdraft, however. Panel b shows directly calculated vertical velocities. Negative vertical velocities of approximately 12 m/s exist well within the cell, inward of the convergence regions. Upward velocities visible to the south are not in cell R and are not of concern here. In panel c, vertical velocities calculated by the upward integration of the mass continuity equation, using \( w = 0 \) m/s at the ground, are displayed. The downdraft calculated in this fashion is weaker, 5 m/s, and is nearer to the convergence zone. Further aloft, at 4.4 km altitude, near the melting level, the engulfment of environmental air is occurring from the south of the cell, as shown in Figure R.10, panel a (as confirmed by tracing air parcel streamlines). The convergence zone appears to be more correlated with the maximum reflectivity than with intersection of the engulfment region with the reflectivity edge, suggesting that precipitation loading, rather than quick cooling of engulfed sub-saturated environmental air, may be an important acceleration mechanism. Panels b and c show the directly calculated and upwardly integrated vertical winds. Both of these calculations show generally upward velocities in the engulfment region. There is some evidence of a downdraft in the directly calculated field. This feature is co-located with the convergence zone in the highest reflectivity.

The same fields presented in Figure R.10 are shown at 2 km altitude in Figure R.11. While there is weak engulfment occurring, the convergence zone is much weaker and is
well correlated with the maximum reflectivity at (52,78). There is much stronger
convergence occurring along a line from (63,72) to (63,85) but this is associated with
cell F. The strong convergence at the southeast corner of the plot area is associated with
cell S. These areas will be discussed in Section 3 which discusses the downdraft in cell S.
(The different characteristics of the convergence regions in the three cells point towards the
possibility of different forcing mechanisms.) The directly calculated vertical velocity field
is shown again in panel b. This is for consistency only, since at this low altitude (2 km),
and this far from the radars, these vertical velocities are not reliable. The vertical velocities
obtained by the upward integration of the mass continuity equation from the ground,
shown in panel c, are accurate at this level and indicate a downdraft of roughly 5 m/s
near (55,75). The situation is similar just below cloud base, at 1 km (800 m above the
ground) as shown in Figure R.12. Panel a shows that engulfment and convergence are
absent. Panel b shows largely meaningless directly calculated vertical velocities, again for
consistency of presentation. Panel c shows the 5 m/s downdraft at roughly (55,75) as
calculated from upward integration of the mass continuity equation.

ii. 2025 Z

As the cell moves towards the radar array, the triple Doppler-analyzed fields become more
reliable. At 2025 Z, the vertical structure of the downdraft is clearly seen. Its upper
extremity is near the melting level at 4.4 km and extends to the divergence region at the
ground. Peak downward air velocities are about 5 m/s.

Figures R.13 through R.16 show vertical slices through the downdraft. Each slice extends
from the ground to 6 km. Radar reflectivity is always contoured to aid in orientation. The
center of the cell is located just to the left of center in all the plots, as indicated by the
45 dBZ contour. In Figure R.13, vertical velocities calculated directly by using triple
Doppler methods are shown. Near the ground these fields are meaningless due to the large
errors in the triple Doppler calculations. Aloft, above about 3 km, they are accurate. The
top of the downdraft is resolved near (80,21), 4.2 km altitude, near the melting level and
the top of the high reflectivity core. The peak resolved downward velocity is about
3-4 m/s. There is an updraft of 13 m/s above the downdraft at (82,28). Figure R.14
shows vertical velocities calculated from vertical integration of mass continuity from the
ground. The velocities are consistent with the directly calculated fields, in the region above
3 km where the direct calculation can be trusted. The origin of the integrated downdraft is
at (80,23), 4.6 km altitude, and there is an updraft of about 9 m/s above it. The peak
downdraft velocity is 5 m/s at 3 km and 1 km altitude. Figure R.15 shows the vertical velocity field calculated by using directly calculated winds above 3.4 km and then integration of mass continuity downward to the ground from 3.4 km. The directly calculated winds at 3.4 km were used as the upper boundary condition. These are the most accurate winds since the direct and integrated techniques are each used in their respective regions of best accuracy. The downdraft has peak negative velocities of about -5.5 m/s at 3 km and at 1.0-1.5 km. The downdraft region is well correlated with the region which exhibits reflectivity greater than 45 dBZ. This has important implications for various forcing mechanisms that will be discussed later in Section f.

The vertical velocity shown at the surface in Figure R.15 is a residual of the downward integration calculation. Deviations from zero indicate that either the boundary conditions were inaccurate or that errors accumulated during the integration. The calculated surface vertical velocities have magnitudes of less than 1.5 m/s almost everywhere in this slice; the average value is -0.2 m/s. This, and the excellent agreement with the upwardly integrated velocities, indicates that the downwardly integrated fields are quite accurate, as discussed in Appendix 4.

Figure R.16 shows horizontal convergence along with cell-relative and slice-parallel wind vectors in a vertical slice through the downdraft. There is a region of inflow to the south (left) of the cell near (80,25) (as confirmed by tracing air parcel streamlines). There is a region of strong convergence, .01 /s, near this region but it is well within the cell at (82,22)-(82,30) and coincident with the highest observed reflectivities. A horizontal slice through the cell, Figure R.17, near the melting level, 4.4 km, shows the flow into the cell from the south and the zone of convergence. There is some convergence near the cloud edge but the major convergence zone is in the cloud interior, well correlated with the highest reflectivity. While in the vertical cross-section it appears that the engulfment is mostly occurring to the south of the cell, this plan view shows that the engulfment may also be occurring from the north. Wind velocities in this northern region are, in fact, roughly parallel to the reflectivity contours as the wind vectors showing upward motion in Figure R.16 suggest.

Triple Doppler analyses directly retrieve precipitation velocities so precipitation fluxes can be calculated. Figure R.18 shows the vertical precipitation particle velocities in the downdraft. The fields below 3 km are unreliable, just as with Figure R.13. The precipitation is moving downwards at up to 13 m/s near 3 km. The precipitation balance
level is just above the top of the air parcel downdraft, at 5 km. Above that, the precipitation is moving upwards at up to 3 m/s. Downward precipitation motion is evident near the melting level. Differential reflectivity, ZDR, data from CP2 were examined in order to distinguish liquid phase precipitation from solid phase precipitation. Figure R.19 shows the ZDR field (shaded) along with precipitation particle vectors. Above 4.5 km, the melting level, the ZDR values are small, < 0.5 dB, indicating that the precipitation was mostly ice. The exception is in the engulfment region where moderate values, near 2.0 dB, are present. The particle vectors indicate that, in the regions of large downward flux, the phase of the precipitation is liquid.

An upper bound on the amount of ice possibly falling into the downdraft can be estimated and calculated. It is unlikely that more ice mass is falling into the top of the downdraft than water mass is falling through the downdraft. Therefore an upper bound on ice mass can be estimated by calculating the precipitation water content in the cell using the Geotis (1971) Z-M relationship. This is shown in Figure R.20. Unfortunately, the different radars measure different reflectivity values in the cell. In the figure, the reflectivity and liquid water content fields using the high reflectivities from CP2 and the lower values from CP4 are shown. Even in the CP2 data, the maximum precipitation water content is only 3.8 g/m³.

Another estimate of ice mass flux can be obtained by calculating the precipitation mass flux through the top of the downdraft and assuming that it is all ice. The ZDR evidence indicates that this is an overestimate, but the measured reflectivity may be low by a factor of about five in regions with significant ice (based on the difference in the dielectric constants of liquid and ice phase water). Therefore, the inferred precipitation water contents from Geotis (1971) may be underestimated. Figure R.21 shows the downward mass flux associated with precipitation. In the region near the melting level, the fluxes are small, around 4 $\frac{g}{m^2s}$. Below this level, much higher fluxes, near $40 \frac{g}{m^2s}$ are observed if the CP2-measured reflectivity is used. A flux of $40 \frac{g}{m^2s}$ would result in a rainrate of 2.4 mm/minute. This is consistent with the 3 mm/minute rate observed under cell F by the surface mesonet station 24 but higher than the 1 mm/minute observed under cell R by station 104. Of course the mesonet stations are unlikely to have experienced direct hits from the most intense regions of the rainshaft. While in a strict sense, the flux value should be used in constructing the upper bound, the high values of ZDR in this region indicate a preponderance of liquid water. The effects of the total precipitation water content
and the ice water content will be discussed later, in Section f, when actual forcing mechanisms are deduced.

Three dimensional divergence serves as a measure of the error in the triple Doppler windfield as discussed in Chapter 2. Since it is expected to be very nearly zero, any deviation is evidence of error, primarily in the vertical windfield calculation. As discussed in Chapter 2, vertical windfield errors grow rapidly near the ground. This is evident in the three-dimensional divergence field displayed in Figure R.22. Squared divergences are very large, exceeding $2.5 \times 10^4$ s$^{-2}$ below 2.2 km, but are generally much smaller elsewhere. Above 3 km, in the downdraft region, three-dimensional divergence is lower than about 2 m/s/gridpoint or .01/s. This indicates that the random errors in the vertical velocities are less than 1 m/s. There are regions of large error near the top of the high reflectivity core and in the engulfment region. Both of these areas exhibit strong reflectivity gradients which are probably the source of the error. Since $w_a$ is calculated from $w_p$ using the Z-w$_t$ formula presented in Pasqualucci (1975) (see Chapter 2), it is prone to large errors in high Z gradient regions. Generally, the three-dimensional divergence field validates the directly calculated fields of Figure R.13 above 3 km altitude.

iii. 2028 Z

Analyses of fields calculated for 2028 Z show that the downdraft is similar in structure to that observed at 2025 Z, but has become slightly more shallow as the cell collapses. Figure R.23 shows the vertical velocity, horizontal divergence, and the differential reflectivity, ZDR, structures in a cross-section through the downdraft. These fields are presented for continuity. Since they are very similar to those at 2025 Z, the discussion here will be brief.

The downdraft is still fairly deep, originating at approximately 3.5 km altitude. (The downward velocities above this level are in a poorly sampled region and are not reliable.) The maximum downward velocities are similar to those at 2025 Z, 6 m/s in the field in which triple Doppler vertical winds are used above 3.0 km and downward integration of the mass continuity equation is used from 3.0 km to the ground (panel b). The zone of horizontal convergence is still in the highest reflectivity at 4-6 km (panel a). The ZDR field (panel c) suggests that the predominant phase of the precipitation at the top of the downdraft is water, although there may be ice at low concentration at the northern edge of the cell, where ZDR values drop below 1 dB within the 45 dBZ total reflectivity contour.
iv. 2031 Z

The time-height history of reflectivity in Figure R.1 shows that the cell R had collapsed considerably by 2031 Z. The surface outflow, which reached a peak near 2028 Z, had also considerably weakened, as had the differential radial velocity, ΔV, across the outflow as shown in Figures R.3 and R.4.

Figure R.24 shows the vertical air velocity and horizontal divergence fields in the downdraft. The triple Doppler $w_a$ field shows that air was moving upwards (slowly, at less than 3 m/s) above 2.0 km. While this directly calculated vertical wind is not very reliable at this low level, it is confirmed by the upwardly integrated $w_a$ field, also shown. The downdraft in this calculation is confined to levels below 2.0 km and has a peak strength below the cloud base at about 600-800 m above the ground. The surface divergence, weaker than at earlier times, is directly under a convergence region located in the highest reflectivity between 1.2 km and 2.4 km altitude. This is much lower than the convergence zone at 2025 Z which was at 5 km altitude. Above the downdraft there is rising motion at about 2 m/s, which is consistent with the directly calculated result. There are weak downward motions at high levels, above 4 km, in the directly calculated vertical wind field. These may not be reliable due to poor radar coverage. However, they are probably indicative of general weak descent associated with the collapse of the cell. Air that is at the 4 km level does not reach the ground (assuming the observed 2 m/s to 3 m/s downward velocity) until well after the study period. The peak reflectivity in the central cross-section of the downdraft has dropped to 44 dBZ. This suggests that precipitation water contents have dropped to near 1 g/m$^3$, if the Geotis (1971) Z-M relation is used, which impacts significantly on the available forcing mechanisms.

The ZDR field, not shown, indicates that in the region of the downdraft (and even well above it) the dominant precipitation type is liquid. Therefore, ice phase precipitation is not influencing the downdraft at this time.

e. Streamline Analyses

Streamlines were calculated and examined in order to measure accelerations following air parcels. Trajectory calculations were attempted, but there was not enough continuity between windfields at different times to inspire confidence in the results.
Figures R.13 through R.15 show the vertical velocity fields in the downdraft at its peak strength. In Figure R.16, the engulfment of air from the south is evident as is the strong convergence at the 4 - 5 km level. Several streamlines with origins in this region were calculated. The traced paths depended on which vertical velocity calculation was used, the direct triple Doppler solution or the dual Doppler integrated solution.

The vertical derivative of vertical velocity, $dw_a/dz$, in the integrated solution must be consistent with the observed horizontal divergence. Therefore, the accelerations calculated using streamlines of this field are dependent on observed horizontal divergence. The dependence is not total because the actual value of $w_a$ determines how rapidly the parcel passes through the $dw_a/dz$ region and thus determines the Lagrangian vertical acceleration increment $dw_a$. A crude calculation indicates that parcels moving with $w_a = -2$ m/s will move through a gridbox volume in about 100 s. Convergence values of 0.5-2.0 m/s/gridpoint are typical in the convergence zone. This suggests that air parcels should accelerate downwards by about 0.5 to 2.0 m/s in 100 s so that parcel acceleration is in the range of -0.005 to -0.02 m/s$^2$.

A typical streamline tracing air entering the cloud from the south is shown below in Table R.1. The parcel starts near the plane of Figures R.13-R.16. (The x grid location of 54 is actually 400 m to the west of the plotted slices.)
Table R.1 Streamline in cell R at 2025 Z. Various properties of an air parcel entering the downdraft are listed. Time in seconds, (x,y,z) grid location in radar grid coordinates, (u,v,w<sub>a</sub>) velocity, reflectivity from the CP4 radar, dBZ, precipitation water content, M, changes in precipitation water content, dM, and changes vertical velocity during listed timestep, dwa.<n>

This streamline was calculated using the raw field of triple Doppler vertical velocity. The parcel accelerates from w<sub>a</sub> = 0 m/s to w<sub>a</sub> = -3.3 m/s during the first 200 s of the trace. As it enters the reflectivity core, it experiences increasing reflectivity and precipitation water content; the reflectivity increases from 42 dBZ to 46 dBZ with a corresponding increase in precipitation water content from 0.8 g/m<sup>3</sup> to 1.6 g/m<sup>3</sup>. While the timestep to timestep vertical accelerations vary greatly, the average acceleration during the first 200 seconds, -0.016 m/s<sup>2</sup>, is more meaningful. This is within the range of values of 0.005 to 0.02 m/s<sup>2</sup> predicted by the horizontal convergence. The consistency of the recorded accelerations with the expected forcing due to thermal buoyancy, precipitation loading, etc., will be discussed below in Section f.

Streamlines calculated using the vertical winds calculated from the vertical integration of the mass continuity equation give similar results. A typical streamline is shown below.
Parcels along this streamline experience similar accelerations to those along the directly calculated streamline. The parcels start with $w_a = -1.5 \, \text{m/s}$ and after 200 s they are falling with $w_a = -4.5 \, \text{m/s}$. The average acceleration is $-0.15 \, \text{m/s}^2$. This is within the range predicted by the horizontal convergence, as discussed above. The reflectivity increases from 41 dBZ to almost 45 dBZ during this period, corresponding to an increase in precipitation water content from 0.7 g/m$^3$ to 1.2 g/m$^3$.

\section*{f. Forcing Mechanisms of the Downdraft in Cell R}

\textit{Simple} calculations of the magnitude of various forcing mechanisms are shown in Section (i). Discussion of the results of a one-dimensional model are presented in Section (ii).

\section*{(i). Calculation of Instantaneous Forcing of the Downdraft in Cell R}

As the air enters cell R, it experiences forcing due to its interaction with the in-cloud environment. As the precipitation water content increases, the loading caused by

\begin{table}
\centering
\begin{tabular}{cccccccccccc}
\hline
Time (sec) & x & y & z & u & v & $w_a$ & dBZ & M & dm & d$w_a$ \\
\hline
2025 Z & \hline
0 & 56 & 81 & 4.2 & 1 & 2.5 & -1.5 & 41 & .7 & --- & --- \\
20 & 56.1 & 81.2 & 4.2 & .6 & 1.8 & -2.6 & 41.5 & .76 & .06 & -1.1 \\
40 & 56.2 & 81.4 & 4.1 & .5 & 1.6 & -2.9 & 41.6 & .78 & .02 & -2 \\
60 & 56.2 & 81.6 & 4.1 & .3 & 1.4 & -3 & 41.8 & .79 & .01 & -1 \\
80 & 56.2 & 81.7 & 4.0 & .01 & 1.1 & -3.3 & 42 & .82 & .03 & -.3 \\
100 & 56.2 & 81.8 & 3.9 & .1 & 1.2 & -3.4 & 42.1 & .83 & .01 & -1 \\
120 & 56.2 & 82 & 3.9 & .04 & 1.2 & -3.5 & 42.1 & .84 & .01 & -1 \\
140 & 56.2 & 82.1 & 3.8 & .7 & 1.8 & -4.2 & 43.4 & 1.03 & .19 & -7 \\
160 & 56.3 & 82.2 & 3.7 & .8 & 2 & -4.3 & 43.7 & 1.08 & .05 & -.1 \\
180 & 56.4 & 82.5 & 3.6 & .8 & 2.2 & -4.4 & 43.9 & 1.12 & .03 & -.1 \\
200 & 56.5 & 82.7 & 3.5 & 1.5 & 1.9 & -4.5 & 44.5 & 1.23 & .11 & -.1 \\
220 & 56.6 & 82.9 & 3.4 & 1.8 & 1.6 & -4.6 & 44.6 & 1.26 & .03 & -.1 \\
240 & 56.8 & 83 & 3.3 & 3.4 & .4 & -4.7 & 46.2 & 1.61 & .36 & -.2 \\
\hline
\end{tabular}
\caption{Streamline in cell R at 2025 Z. Various properties of an air parcel entering the downdraft are listed. Time in seconds, (x,y,z) grid location in radar grid coordinates, (u,v,$w_a$) velocity, reflectivity from the CP4 radar, dBZ, precipitation water content, M, changes in precipitation water content, dm, and changes vertical velocity during listed timestep, d$w_a$.}
\end{table}
precipitation increases. Assuming an air pressure of about 600 mb, typical of the 4.5 km level at the top of the downdraft, and a temperature of roughly 277 K, the air density is about 750 g/m³. The acceleration caused by precipitation loading is $g \frac{\rho_{\text{prec}}}{\rho_{\text{air}}}$. The following Table, R.3 shows accelerations that would be caused by various amounts of liquid water content, as measured by radar, at different levels in the atmosphere.

<table>
<thead>
<tr>
<th>Reflectivity dBZ</th>
<th>$\rho_{\text{prec}}$ (g/m³)</th>
<th>600 mb 277 K</th>
<th>800 mb 289 K</th>
<th>1000 mb 303 K</th>
</tr>
</thead>
<tbody>
<tr>
<td>35</td>
<td>.27</td>
<td>.0035</td>
<td>.0027</td>
<td>.0023</td>
</tr>
<tr>
<td>40</td>
<td>.60</td>
<td>.0078</td>
<td>.0061</td>
<td>.0051</td>
</tr>
<tr>
<td>45</td>
<td>1.33</td>
<td>.017</td>
<td>.014</td>
<td>.011</td>
</tr>
<tr>
<td>48</td>
<td>2.16</td>
<td>.028</td>
<td>.022</td>
<td>.018</td>
</tr>
<tr>
<td>50</td>
<td>2.98</td>
<td>.039</td>
<td>.030</td>
<td>.025</td>
</tr>
<tr>
<td>52</td>
<td>4.11</td>
<td>.053</td>
<td>.042</td>
<td>.035</td>
</tr>
<tr>
<td>54</td>
<td>5.67</td>
<td>.074</td>
<td>.058</td>
<td>.048</td>
</tr>
<tr>
<td>56</td>
<td>7.83</td>
<td>.10</td>
<td>.080</td>
<td>.067</td>
</tr>
<tr>
<td>58</td>
<td>10.80</td>
<td>.14</td>
<td>.11</td>
<td>.092</td>
</tr>
<tr>
<td>60</td>
<td>14.90</td>
<td>.19</td>
<td>.15</td>
<td>.13</td>
</tr>
</tbody>
</table>

Table R.3 Downward acceleration caused by precipitation loading. The acceleration produced by various precipitation water contents, as measured by radar reflectivity, at three different pressure levels. Temperatures are those typical at the respective levels. Units of acceleration are m/s². The Geotis (1971) Z-M relation is used.

The observed reflectivity (CP4 radar) along the two presented streamlines was 41–46 dBZ. According to the table, this should result in downward accelerations of about .01-.02 m/s² which is very close to the observed values of -.015 m/s².

While the observed acceleration would seem to be explained, other mechanisms may have been acting on the parcel. Since the parcels are observed to be entering the cloud (or at least the detectable radar edge) from the outside environment, subsaturated air is being brought into contact and possibly mixed with cloudy air containing raindrops. Both the cloud droplets and raindrops will evaporate and cool the parcels, causing accelerations due to negative buoyancy.

This forcing, assuming a very simplistic scenario where parcels are held stationary, can be calculated. In this scenario, subsaturated environmental air is engulfed into the precipitation core. It does not mix with the cloudy air, but precipitation falls through it and evaporates, producing cooling. The parcel will eventually cool to its wet bulb temperature,
$T_w$, and the forcing can be calculated from the post evaporation virtual temperature. Using values from the 18 Z Redstone sounding $T_w(\text{environment}) = 272.9$ K and $T_v(\text{environment}) = 277.8$ K. Thus,

$$\text{downward acceleration} = g \frac{(T_v(\text{environment}) - T_v(\text{environment}))}{T_v(\text{environment})}$$

$$= 9.81 \text{ m/s}^2 \frac{(277.8 \text{ K} - 274.0 \text{ K})}{277.8 \text{ K}} = .134 \text{ m/s}^2.$$

This acceleration is very large and is not observed in the downdrafts. The assumption that raindrop evaporation alone can cause a rapid instantaneous cooling is very flawed. Raindrop evaporation, discussed in Appendix A.2, is a slow process. An environmental parcel representative of that observed on the study day would require many minutes to saturate and cool to its wet bulb temperature. Therefore, the full negative buoyancy would not be instantly realized. Table R.4 illustrates the time scales involved in this process.

<table>
<thead>
<tr>
<th>Time (s)</th>
<th>T (K)</th>
<th>$T_v$ (K)</th>
<th>accel (m/s²)</th>
</tr>
</thead>
<tbody>
<tr>
<td>0</td>
<td>277</td>
<td>277.8</td>
<td>.000</td>
</tr>
<tr>
<td>60</td>
<td>276.6</td>
<td>277.4</td>
<td>.014</td>
</tr>
<tr>
<td>120</td>
<td>276.2</td>
<td>277.0</td>
<td>.028</td>
</tr>
<tr>
<td>300</td>
<td>275.3</td>
<td>276.2</td>
<td>.056</td>
</tr>
<tr>
<td>600</td>
<td>274.3</td>
<td>275.3</td>
<td>.088</td>
</tr>
<tr>
<td>900</td>
<td>273.8</td>
<td>274.8</td>
<td>.106</td>
</tr>
</tbody>
</table>

Table R.4 Cooling and acceleration caused solely by the evaporation of raindrops. In this calculation the air parcel has been held stationary and so no adiabatic compressional effects are included. A reflectivity of 45 dBZ has been used along with an environmental virtual temperature of 277.8 K and a pressure of 600 mb.

The data in Table R.4 show the rate at which an air parcel would cool if raindrops with a concentration corresponding to a radar reflectivity of 45 dBZ were to fall through it. The values have been calculated using the evaporation scheme discussed in Appendix 2. Even after 600 to 900 seconds, the parcel has not cooled to its wet bulb temperature of 272.9 K. The acceleration that such a parcel would feel, as a function of time, is also listed. During the first 120 seconds, the accelerations are consistent with the observed accelerations along the streamlines. The parcel, in this simplistic calculation, continues to cool and become increasingly negatively buoyant. In reality, the parcel would fall during this time, compress adiabatically and warm. In addition, it might mix with in-cloud air which is thermally buoyant and contains cloud droplets.
While differential reflectivity measurements show little evidence for significant ice concentrations in cell R, it is useful to calculate the possible cooling that would result if ice were to fall into the parcel and melt. Again, some crude assumptions must be made. The equivalent of the $T_w$ assumption above is to assume that the parcel will be brought to 273 K by the melting. While this is unrealistic at lower levels, it is plausible near the melting layer where temperatures are very close to 273 K anyway. It is immediately obvious that this would produce large negative buoyancies and downward accelerations since 273 K is near the observed value of $T_w$. As was seen to be the case with raindrop evaporation, the melting rate may be so slow as to make this assumption unreasonable. By picking reasonable but generous numbers for ice particle terminal velocity, ice particle density, and melting layer thickness, a crude estimate of the cooling rate can be calculated. Srivastava (1987) shows that melting ice particles over 1 mm in diameter require over 1 km to convert 1/2 of their mass to liquid, so a melting layer thickness of 1 km is assumed.

The cooling rate is:

$$\frac{dT}{dt} = \frac{w_t \rho_i L}{Z_{ML} C_p \rho_a} = .007 \text{ K/s} = 0.4 \text{ K/minute},$$

where $w_t = 8 \text{ m/s}$ is the ice particle terminal velocity, $\rho_i = 2 \text{ g/m}^3$ is the ice particle density, $\rho_a = 750 \text{ g/m}^3$ is the air parcel density, $Z_{ML} = 1000 \text{ m}$ is the thickness of the melting layer, $L$ is the latent heat of fusion, and $C_p = 1005 \frac{J}{\text{kg K}}$ is the specific heat of air. After one minute this cooling could provide a downwards acceleration of .014 m/s$^2$, which is comparable to that observed at the top of the downdraft. The term in the numerator in the above equation, $w_t \rho_i = 16 \frac{g}{\text{m}^2\text{s}}$, is the mass flux into a moving air parcel. This value is less than the worst case mass flux of 40 $\frac{g}{\text{m}^2\text{s}}$ mentioned in Section d, but the latter value is the flux through a fixed surface, not flux through a descending parcel.

ii.. Model Simulations of the Downdraft

Since evaporation of raindrops, evaporation of cloud droplets, precipitation loading, melting of ice particles, adiabatic compression, mixing between the environment and the
cloud and even small effects such as conductive cooling by raindrops and ice particles all may affect the air parcels in the downdraft, the simple plausibility calculations presented above are not adequate to determine the dominant forcing mechanisms that are present. The one-dimensional model used to diagnose the relative strength of these factors is described in Appendix 2. The model assumes that the downdraft is an entraining plume and the entrainment environment can be specified along with other entrainment parameters.

Two model simulations are presented below in Table R.5 and Table R.6. The primary difference in the two realizations is the type of the air that is entrained into the downdraft as it descends. In the first simulation, the air is entrained from the out-of-cloud environment. The environment contains dry, low $\theta_e$ air that is potentially negatively buoyant. As the air is mixed into the downdraft it pulls the downdraft air towards sub-saturation. The downdraft is cooled as raindrops fall through this subsaturated air. In the second simulation cloudy air is entrained into the downdraft. The cloud contains cloudy, relatively warm, saturated, high $\theta_e$ air that is generally positively buoyant. As the cloudy air is mixed into the downdraft, it tends to warm it but it also supplies cloud droplets which can rapidly evaporate and help maintain almost wet-adiabatic descent. True wet-adiabatic descent cannot be achieved because, under realistic assumptions, insufficient cloud droplets are entrained to maintain saturation. A simple schematic representation of the two processes is shown in Figure R.25.

Both presented model simulations were initialized with:

- Radar reflectivity $\text{dBZ} = 50$ dBZ
- Cloud droplet concentration in initial parcel $\rho_y = 0$ g/m$^3$
- Ice particle concentration in cloud $\rho_I = 0$ g/m$^3$
- Entrainment parameter $\alpha = .25$
- Initial radius $r = 500$ m
- Pressure level of simulated boundary level $\text{PBL} = 960$ mb
- Pressure level of ground $\text{PGR} = 1000$ mb
- Cloud droplet concentration in cloudy air $\rho_{yc} = 1.0$ g/m$^3$

The model soundings were: in (pressure, temperature, relative humidity) format:

<table>
<thead>
<tr>
<th>Environment</th>
<th>Pressure (mb)</th>
<th>Temperature (°C)</th>
<th>Relative Humidity (%)</th>
</tr>
</thead>
<tbody>
<tr>
<td>outside environment</td>
<td>(600,277,.55)</td>
<td>(800,289,.90)</td>
<td>(1000,304,.65)</td>
</tr>
<tr>
<td>outside entrainment environment</td>
<td>(600,277,.55)</td>
<td>(900,295,.75)</td>
<td>(1000,298,.95)</td>
</tr>
<tr>
<td>cloudy entrainment environment</td>
<td>(600,278,1.)</td>
<td>(900,294,1.)</td>
<td>(1000,297,1.)</td>
</tr>
</tbody>
</table>
Through most of their depth, both model downdrafts have downward velocities of about 3 m/s to 5 m/s. This is very close to the radar observations shown in Figures R.14 and R.15. The model downdrafts both increase in strength at the lowest levels, below 900 mb. This is not consistent with the radar observations. The real and model atmospheres are almost neutrally stratified near the ground. Therefore, cool descending parcels feel large accelerations. Part of the greater strength in the model downdrafts is due to the shallowness of the model boundary layer. The model does not attempt to accurately simulate this boundary layer and it is typically thinner than the radar observations suggest. The observed surface divergence layer starts at about 600 m to 800 m above the surface, whereas the model boundary level begins at 400 m above the surface. Thus, downdrafts continue to accelerate at low levels in the model whereas they decelerate in the real atmospheric boundary layer.

The thermodynamic qualities of the model downdrafts compare favorably with the measurements taken at the mesonet stations presented above in Section c. Both simulations

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produce surface outflow air that has significantly reduced $\theta_e$. The outside entrainment run produces a value, 342 K, that is very near the lowest observed at the surface, 340 K, (see Figure R.7). The cloudy entrainment run produces a value that, while still significantly depressed when compared to the surface air in the model, is somewhat higher, 349 K. In both simulations, downdraft parcels arrive at the ground both warmer and drier than the mesonet observations suggest. The outflow observations average about 297 K in temperature and 19 g/m$^3$ in vapor density, while the model predicts about 301-3 K temperature and 14.5-18.0 g/m$^3$ vapor density, depending on whether in-cloud or out-of-cloud entrainment is used. The differences are due to two factors. First, in both the real and model boundary layers, parcels are slowed and therefore have more time to saturate and cool as rain falls through them. Since the model boundary layer is too thin, less slowing occurs and thus there is less moistening and cooling. Second, the observed air may have travelled along the ground for a minute or more before impinging on a mesonet station.

The outside entrainment and cloudy entrainment simulations represent extremes of the mixing environment parameter space. Streamline observations from radar suggest that some mix of cloudy and environmental air is entrained into the downdraft during its descent. Despite deviations of the model simulations from the radar and surface observations (the dryness of the outflow and the overestimate of low-level downdraft strength), both simulations reproduce the changes in $\theta_e$ and $\rho_v$ observed at the surface as well as the vertical velocities observed by radar fairly closely. The in-cloud entrainment simulation results in surface $\theta_e$ values that are too high and the out-of-cloud simulation produces values that are somewhat lower than the observations at most mesonet stations. Simulations that use a parameterized mix of cloudy and outside air produce intermediate results; it is this mix of in-cloud and out-of-cloud air that most accurately duplicates both the surface observations and the radar observations.

In both simulations, the dominant forcing mechanism appears to be the loading caused by precipitation, which causes accelerations that average about .03-.04 m/s$^2$. This result is sensitive to the reflectivity values used; a reduction of just 3 dB, which is comparable to the differences in reflectivities measured by the various Huntsville radars, would reduce the loading forcing term by 30 percent according to the Geotis (1971) formulation. The acceleration due to forcing in the simulations was higher than that predicted in the simple calculations of Section (i) concerning the topmost portion of the downdraft because the reflectivities in the model were representative of the bulk of the downdraft, not the
somewhat lower values observed in its uppermost regions. As precipitation evaporates and air density increases in the model downdraft, the loading decreases and therefore the loading forcing is smaller near the ground, \(0.01 \text{ m/s}^2\) at 950 mb.

The outside entrainment simulation produces a downdraft that is upwardly thermally buoyant above 850 mb. The cloudy entrainment simulation produces a downdraft that is almost neutrally buoyant above 750 mb and upwardly buoyant from 750 mb to 850 mb. Below 850 mb where, as mentioned above, the environment is more unstable, both simulated downdrafts are strongly negatively buoyant. Near the ground, 950 mb, the forcing due to thermal buoyancy, \(0.03 \text{ m/s}^2\), is two to three times larger than that due to precipitation loading.

While the small values of the thermal buoyancy acceleration terms appear to suggest that raindrop evaporation is not an important forcing mechanism at higher altitudes, this conclusion would be incorrect. Significant evaporation is required in order to keep the descending parcels cool enough to remain neutrally, or only slightly positively, buoyant. In model runs with no evaporation, the descending parcels rapidly warm so that precipitation loading is not strong enough to cause further descent. During the entire summer of 1986 in Huntsville, no outflows with increased temperature at the surface were observed.

The role of cloud water is important in the simulation in which cloudy air is entrained into the downdraft. In this downdraft, 40 percent (4.6 K of 11.7 K), of the cooling caused by evaporation was due to cloud droplets. When cloudy air is entrained the cloud droplets evaporate much more rapidly than the raindrops (which evaporate very slowly, see Table R.4, and Appendix 2) and allow the downdraft to absorb much more water. Thus, the cloudy entrainment downdraft has \(\rho_c(800 \text{ mb}) = 11.2 \text{ g/m}^3\), which is 40 percent higher than the dry entrainment simulation. The extra cooling provided by these drops is enough to counter the upwardly buoyant effect of entraining the warm in-cloud air.

The role of melting ice is explored with the model and found to be insignificant. The model is initialized with 2 g/m\(^3\) ice, representing 2/3 of the total water content. The resultant downdrafts are similar to the cases without ice. Melting, occurring mostly in the first 500 m of the downdraft, cause about 2.5 K of cooling. The downdrafts are somewhat stronger (\(w_a = -7 \text{ m/s}\)) at upper levels than both the model cases without ice and the observations. The thermodynamic qualities of the air at the ground are very similar to the
simulations without ice. The simulated outflow air was slightly drier, by 0.2 g/m\textsuperscript{3}, and exhibited a slightly lower, by 0.6 K, value of $\theta_e$. In the model simulations, the cooling by the melting of ice tends to substitute for, and not add to, the cooling due to evaporation of raindrops. This is because the downdraft air is cooler at upper levels and is consequently less able to evaporate the raindrops falling through it, and because water vapor deposits on the cold water and ice mass in the upper reaches of the downdraft.

g. Summary

Multiple Doppler analyses have shown that the downdraft in cell R has origins near the melting layer at 4-5 km altitude near the $\theta_e$ minimum in the proximity sounding shown in Figure 1.3. Streamline analyses indicate that outside air is entrained into the top of the descending precipitation core which contains little or no ice at altitudes where downward precipitation flux is occurring. Accelerations calculated along streamlines are consistent with the radar observed convergence and forcing fields. The observed downdraft penetrates to the ground and contains maximum downward vertical velocities of about 5-6 m/s at 1.5 and 3 km altitude.

At the surface, an outflow is observed by both radar and several mesonet stations. The thermodynamic characteristics of the outflow air are very different from that of a nearby outflow from cell F. The outflow air from cell R has significantly reduced $\theta_e$ and $\rho_v$, the latter being especially noteworthy since it occurs during heavy rain, whereas the outflow from cell F exhibits nearly constant $\theta_e$ and increased $\rho_v$, as will be discussed in Section 2.

Simple model simulations reproduce the bimodal velocity maximum at 1.5 and 3 km and the low surface $\theta_e$ and $\rho_v$. The model simulations suggest that precipitation loading and evaporation of raindrops are the dominant forcing mechanisms in this downdraft, although the evaporation of cloud droplets is necessary if warm cloudy air is entrained into the downdraft during descent. The role of ice is not supported by the observations and the model simulations that include ice produce a somewhat less realistic downdraft.
2. Cell F

a. Introduction

This Section describes the structure and physics of the downdraft in the cell that caused the outflow labelled F in Figure 3.C.1. The downdraft had origins at low-levels, about 2 km altitude, lower than the downdraft in cell R documented earlier in Section 1. It was forced primarily by precipitation loading and evaporation of raindrops.

b. Cell F history

The time history of reflectivity in this cell is shown in Figure F.1. The cell develops slowly after about 2016 Z. The 50 dBZ contour passes the melting level at that time and only rises to about 6 km by 2031 Z. There is no evidence of storm collapse during the period that the cell is under observation by radar. Unfortunately, after about 2034 Z, multiple Doppler analysis was impossible because the radars did not scan appropriately. Peak reflectivities below the melting layer are higher than in the nearby cell R, reaching about 55 dBZ through most of the region beneath the melting level by 2031 Z.

Figure F.2 shows the time history of surface divergence along with contours of reflectivity at the 800 m level. The large scale structure shows convergence at the leading edge of the storm region; air is rising from the surface and, as will be seen in this Section, entering the downdraft and descending to the ground. This large scale structure appears very similar to the horizontal role downdraft, MB1, described in Mahoney and Elmore (1991).

Weak divergence exists under the cell at about (65,80) at 2019 Z. At this time it is difficult to identify the reflectivity or divergence signature of the cell as distinct from its neighbors. The surface divergence pattern remains weak until about 2025 Z. From 2025 Z until 2034 Z, peak divergence values of .02 /s exist continuously near (60,85) to (60,90). This is equivalent to 20 m/s per kilometer, a respectable value. The magnitude of the divergence increases only slightly from 2028 Z to 2034 Z. The peak values occur in a region that is well correlated with the peak reflectivities observed at 800 m altitude. Peak reflectivities near 55 dBZ exist during this time, values that are higher than during the period of low divergence (2019 Z to 2025 Z). The more standard single Doppler measure of outflow strength, differential radial velocity, ΔV, shows a history that is consistent with the multiple Doppler horizontal divergence results, as seen in Figure F.3. ΔV remains small from...
2013 Z until 2022 Z. The actual value of $AV$ was difficult to determine at these times because of the interfering outflows from nearby cells. After 2025 Z, however, the outflow was distinct and strong, with $AV$ climbing to and remaining at 17 m/s.

c. Surface Mesonet Observations Below Cell F

Cell F preceded cell R through the Huntsville surface mesonet array. Figure F.4 shows the path of the outflow area, as marked by the surface divergence. The radar grid-relative coordinates are labelled and shift as described in the discussion of cell R. The outflow crosses several mesonet stations, including stations 104 and P24. The time series from these stations, presented in Figures R.7 and R.8 (in Section 1, above), show the effects of both cell F and cell R.

Cell F affects station 104 from about 2023 Z until 2028 Z. At the beginning of this period, the windspeed increases from 4 m/s to 12 m/s, the temperature drops from 304 K to 298 K, and the relative humidity increases from .50 to .90. Importantly, in light of the model comparisons that will follow in a later Section, $p_v$ increases from below 18 g/m$^3$ to 21 g/m$^3$. $\theta_e$ increases from a pre-storm value of 349 K (somewhat low compared to neighboring stations) to 353 K during the event. At least some of this increase is due to the $\theta_e$ jump effect described in Appendix 1. The $\theta_e$ and $p_v$ behavior during the passage of cell F can be contrasted with that which occurs during the passage of cell R. In the latter case, the values of both quantities drop sharply. The distinct and characteristic behavior during the passage of cell F will figure importantly when the vertical structure of the downdraft is discussed in Section d, below.

Station 24 is also affected by cell F and the general behavior is quite similar. The outflow first impinges on the station at 2025 Z, somewhat earlier than the radar-derived surface divergence would indicate. It is likely that the surface divergence field was calculated from data that were taken mostly around 2023 Z or 2024 Z and included in the nominal 2025 Z grid. As the outflow hit the station, the wind increased from 3 m/s to 10 m/s, the temperature decreased from 303 K to 297 K, and the relative humidity increased from .55 to .90. Vapor density, $p_v$, increased from 17.5-18.0 g/m$^3$ to 20.5 g/m$^3$. $\theta_e$ remained relatively constant, increasing only slightly from 351 K to 353 K. While no rainfall was observed several minutes earlier, when the event crossed station 104, heavy rain, at rates up to 3 mm/minute, fell at station 24 from 2028 Z until 2030 Z. This heavy rain was occurring under the 55 dBZ echoes noted above in the discussion of Figure F.2.
The surface observations of nearly constant \( \theta_c \) and increased \( \rho_v \) will have to be explained by examining the vertical structure of the downdraft using radar evidence and when using the constrained model to diagnose forcing mechanisms.

d. History of Cell F as Deduced from Triple Doppler Analyses

Radar analyses similar to those conducted on cell R were also done for cell F. Somewhat less detail will be presented here since the techniques have been elucidated in the Section concerned with cell R. The cell was observed continuously from about 2022 Z to 2031 Z, during which time it crossed into the radar array, permitting detailed triple Doppler analyses.

i. 2022 Z

As is evident in the divergence history illustrated in Figure F.2, a weak outflow was already in progress at 2022 Z. Vertical cross-sections through the downdraft show that it has fundamentally different characteristics than cell R, being confined to levels below 2 km altitude.

Figures F.5 through F.10 are north-south cross-sections extending from the ground to 4 km altitude. Figure F.5 shows the directly calculated vertical velocity field (after some smoothing to eliminate features with wavelengths less than 400 m). While the 2022 Z vertical velocity field in the cell R region was highly inaccurate, the proximity of cell F to the radar array (though still not inside it) provides some hope that it may be better. In order to interpret this field and evaluate its reliability, it is useful to refer to Figure F.6, in which the number of observing radars in each region of the plot area is shaded. The region at the upper left corner of the plot area, as well as scattered gridpoints near the ground, were sampled by only three radars. In fact, the three radars that sampled in that region had an unfavorable geometry and the vertical velocities are therefore quite suspect. \( \nabla \cdot V \) calculations confirm this, as shown in Figure F.7. \( \nabla \cdot V \) is very high in the upper left, three radar, region of the plot. (The region of small \( \nabla \cdot V \) at the extreme upper left portion of the plot area is actually a region where vertical velocities were of such poor quality that they were assigned to undetermined values by the analyses software and the \( \nabla \cdot V \) algorithm set \( \frac{\partial w_v}{\partial z} = 0 \). Since the data in this region were extremely poor, nothing would have been gained by correcting this software limitation.) The region sampled by all five radars is
confined to a thin swath from just above the ground to between 1.0 km and 1.6 km altitude. In and near this region, suitably far from the ground, $\nabla \cdot \mathbf{V}$ values are at a minimum, suggesting accurate values of $w_a$ have been calculated. The fifth observing radar in this case is UND, a radar very near to the cell; its observations were taken at fairly high elevation angles compared to the other radars. The error reducing effect of greater sampling overcomes the problems usually associated with low altitude. The expected vertical velocity variances are lower in the five radar region than in the region just above which is sampled by only the four more distant radars.

With all this in mind, a meaningful interpretation of the vertical velocity field in Figure F.5 is possible. The figure shows a shallow downdraft confined to levels below 1.6 km altitude (recall that cell R extended to 4 km altitude). The downdraft appears to be deeper at its northern edge, but this is in the region of less accurate calculation. Near the ground the values are positive (upward), but direct calculations near the ground are highly inaccurate and therefore meaningless. The peak downdraft velocities are -5 m/s in the region sampled by five radars and -9 m/s in the region sampled by only four radars.

Vertical velocities calculated from vertical integration of the mass continuity equation are likely to be more reliable than the direct triple Doppler results at the 2022 Z storm location, far from the radars. This is particularly true in the downdraft region close to the ground. Figure F.8 shows integrated vertical velocities in the downdraft region. The field confirms the general pattern noted above in Figure F.5. There is a shallow downdraft, confined below 1.8 km altitude, with a peak strength of -5 m/s. Above 2 km there is an updraft, except possibly at the northern edge of the cell near $(y,z) = (75,15)$, in the region where the direct calculation indicated a downdraft. In this region the vertical velocities are close to zero.

Figure F.9 shows the horizontal divergence field. The strong ground level divergence reaching .02/s in the outflow is visible at (67,1). The most notable feature of the field is the strong convergence from 1 km to 3 km. This is in contrast to the mid level, 4 km to 5 km, convergence zones characteristic of cell R. (The extension, here, of the convergence zone to 4 km is suspect because this is in the region of poor radar sampling in which errors in horizontal wind calculations are large.) There is a region of convergence at 4 km at the extreme northern edge of the cell that may be real but, as the cell-relative wind vectors of Figure F.10 show, only air below 2 km is being drawn downward to the ground. This convergence is associated with another cell to the north. The region of
convergence associated with the high reflectivity region to the south at (60,7) is part of cell S and the air in that cell is moving upwards and not into the downdraft of cell F. The wind vectors show that slice-parallel horizontal winds are weak, generally about 2 m/s from the north.

A horizontal slice through the cell at the altitude of the maximum horizontal convergence, 1.8 km, shows that air is being engulfed into the cloud from the front and possibly the rear. Figure F.11 shows both the convergence field and the horizontal wind field. The convergence, with peak values near -.015/s, is well within the cloud and is well correlated with the maximum reflectivity. The edge of the radar-detectable reflectivity is off the right edge of the plot and is thus at least 2 km to 3 km from the convergence zone (the true cloud edge is even further). The cell-relative horizontal windfield vectors show that air is being engulfed into the cloud and into the convergence region from the right (east) at about 7 m/s. As streamline analyses will confirm (Section e), air is being entrained into the cloud from the leading side of the cell, entering the high reflectivity region and then being accelerated downwards towards the ground into the outflow.

The location of the convergence zone at 1 km to 3 km altitude indicates that sub-cloud evaporation of raindrops is probably not a major forcing mechanism at the top of this downdraft. As noted in Section B, the lifted condensation level was near 900 mb, or about 1 km (see Figure 3.B.2).

The correlation of convergence and high reflectivity suggests that, as with cell R, precipitation loading may be a major forcing mechanism in this downdraft. In addition, the displacement of the convergence zone from the cell edge by at least 3 km indicates that rapid cooling of sub-saturated entrained air is not a major factor. A detailed examination of the relative importance of various mechanisms will involve the model study presented below in Section f.

ii. 2025 Z

Three minutes later, at 2025 Z, the picture is very much the same. A downdraft exists below 2.2 km, as shown in Figure F.12, and the peak downdraft velocity is about -4.0 m/s. Convergence is occurring in a broad zone from 1.2 km to 3.0-3.6 km altitude in the maximum reflectivity, well away from the cloud edge. While the convergence was occurring at altitudes up to 3.6 km, the air entering the downdraft had origins below
2.0 km. Above the downdraft, at altitudes extending from 2 km to at least 4 km altitude (the highest level displayed), there is a strong updraft with vertical velocities up to 11 m/s at 4 km.

iii. 2028 Z

The downdraft continued at 2028 Z. Figure F.13 shows the same parameters as the previous figure, but at 2028 Z. The downdraft is very similar in structure, though slightly more intense. The peak downward velocities are about -5 m/s to -6 m/s at about 1 km altitude and the downdraft depth is about 2 km. The convergence zone still extends from 1 km to 3 km altitude in the highest reflectivity region. As can be seen, an extensive area of 50 dBZ echo exists from just above the ground upwards to about 3.8 km. (The lower reflectivities near the ground are due to beam clipping by ground obstruction and smoothing prior to contouring.) This reflectivity corresponds to precipitation water contents of 3 g/m³ or higher through a broad region, using the Geotis (1971) Z-M formulation. Peak observed reflectivities of 55-56 dBZ are indicative of about 7 g/m³ precipitation water content. Therefore, large amounts of water are available for both loading and evaporation (and, in theory, melting, but the downdraft is confined to levels that exhibit no evidence of ice as indicated by differential reflectivity measurements, not shown).

iv. 2031 Z

The outflow was at nearly peak strength at 2031 Z. By this time the cell had moved well into the interior of the radar array, potentially reducing the errors in the directly retrieved vertical velocities. At 2022 Z, vertical velocities at the top of the downdraft were just resolvable because five radars, including a very proximate one, UND, sampled the region. At 2025 Z and 2028 Z, only four radars took data in this cell. Therefore, the directly retrieved fields at the top of the downdraft were too inaccurate to be useful. The cross-sections presented at those times show only the upwardly integrated vertical winds. At 2031 Z, while most of the cell was sampled only four times, the cell-array geometry, and the possible assistance of a fortunate combination of sampling times (minimizing evolution errors), allowed direct calculation of vertical velocities down to about 1.6 km.

Figures F.14 through F.18 show calculated fields in a vertical slice through the downdraft extending from the ground to 4 km altitude. Figure F.14 shows the number of radars that
sampled various regions in the slice. Through most of the reflectivity core and, as can be seen by comparing with Figure F.16 and Figure F.17, through most of the downdraft, this is four radars. At the extreme left (southern) edge, five radars took data; only three sampled the rightmost region. The measured \((\nabla \cdot V)^2\) shows the effect of this observation pattern. Since \((\nabla \cdot V)^2\) is a measure of triple Doppler error, the values should be lower in more frequently sampled regions. In Figure F.15, the strong altitude dependence is obvious with values of 20 (m/s/gridpoint)\(^2\) = 0.0005/s\(^2\) typical near the ground. Near the ground in the region sampled by five radars, the values are lower, 0.0003/s\(^2\). \((\nabla \cdot V)^2\) is generally much smaller above 1.6 km (values frequently below 4 x 10\(^{-5}\)/s\(^2\)) and it is at this level that the boundary condition between directly calculated and downwardly integrated vertical velocities is applied.

Between 2 km and 4 km, in the region sampled by five radars, \((\nabla \cdot V)^2\) is less than 1 (m/s/gridpoint)\(^2\) or 2 x 10\(^{-5}\)/s\(^2\). Figure F.16 shows the directly calculated vertical velocities in the region of low \((\nabla \cdot V)^2\) and the combined direct and integrated vertical velocity field (with slight smoothing). In the raw field the roots of the downdraft can be seen near the 2 km level, just above the level where the velocities become unreliable. In the combined field, the full downdraft can be seen. The peak downward velocity is about -9 m/s at 1 km above the surface. The downward integration ends at the ground with a residual of about -1.5 m/s downward velocity, where, ideally, 0 m/s should be the result. This indicates that the peak velocities at 1 km are probably overestimated by about .8 m/s and were actually near -8 m/s. Velocities from upward integration, Figure F.17, confirm the basic structure of the downdraft with a 2 km depth and a maximum downward velocity of -7 m/s. Figure F.18 shows strong horizontal convergence in the broad zone from 1.4 km to 3.4 km, peaking near 2 km altitude.

The pattern of engulfment into the highest reflectivity region coincident with the maximum horizontal convergence has also persisted from 2022 Z, as seen in the 2 km slice through the cell presented in Figure F.19. Inward, mostly easterly, velocities of about 7 m/s towards the horizontal convergence zone are evident. This zone is extremely well correlated with the 50 dBZ contour which surrounds peak reflectivities of nearly 55 dBZ.

e. Streamline Analyses

Streamlines were calculated and examined in order to measure accelerations following air parcels. These streamlines were calculated by assuming that the velocity and reflectivity

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fields did not evolve during the tracing period and that the only changes in those fields were due to storm motion. Comparisons of the presented reflectivity and velocity fields shown in Section d show that this assumption is flawed. In fact, evolution of the fields was large enough so that the streamlines made with fields from any particular grid time could not be matched with grids from the times adjacent to it. Therefore, trajectory calculations were not possible. A typical streamline tracing air entering the downdraft is shown below in Table F.1. The streamline is traced forwards and backwards from the 2 km level.

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Table F.1 Streamline entering downdraft in cell F at 2031 Z. Various properties of an air parcel entering the downdraft are listed. Time in seconds, (x,y) grid location, z km, (u,v,w_a) velocity, reflectivity from the CP4 radar, precipitation water content, M, changes in precipitation water content and vertical velocity during listed timestep, dM and dw_a.

This streamline is calculated using the raw triple Doppler vertical velocity field. The parcel enters the highest reflectivity region and is accelerated downwards. The parcel experiences downward accelerations of approximately .04 m/s² while accelerating in the > 50 dBZ
region. In this region the precipitation water content exceeds 3 g/m³. Peak downward velocities are nearly -9 m/s. As the parcel nears the ground, it decelerates to -3 m/s. Horizontal convergence observed in the uppermost regions of the downdraft (see Figure F.19, for example) is about -2.0 m/s per gridpoint, -.01 /s, which is consistent with -.02 m/s² accelerations (see discussion in cell R, Section e).

f. Forcing Mechanisms of the Downdraft in Cell F

Simple calculations of the magnitude of various forcing mechanisms in the downdraft of cell F are shown in Section (i). Discussion of the results of a one-dimensional model are presented in Section (ii). The one-dimensional model is more thoroughly discussed in Appendix 2.

i. Calculation of the Instantaneous Forcing of the Downdraft in Cell F

Unlike cell R, both the region that exhibits horizontal convergence and the region in which air parcels are accelerated downwards are well within the radar reflectivity boundary of the cell. The cell's reflectivity edge is several minutes upstream from this region. Evaporation calculations show that unsaturated parcels would become nearly saturated during the time that it would take them to travel this distance. In fact, the true travel distance is even greater than indicated in the horizontal slices above since the true cloud edge is probably outside the reflectivity edge. Therefore, instantaneous cooling by the entrainment of subsaturated air is not a likely acceleration mechanism in this downdraft.

The top of the downdraft, at about 2 km, lies well below the melting level in a predominantly liquid water environment. Melting of ice is therefore not implicated as an acceleration mechanism in the downdraft in cell F. Therefore, the only instantaneous downward forcing term that is relevant in this case is precipitation loading. Following the calculations shown in Table R.3 (in Section f of the cell R discussion), for reflectivities of 50 dBZ - 54 dBZ and pressure levels between 800 mb (2 km) and 1000 mb (surface), the expected accelerations in the downdraft are .03 m/s² - .04 m/s². This is consistent with the observed value of .04 m/s².
ii.. Model Results

The one-dimensional model was run in order to simulate the cell F downdraft. The model was initialized using saturated, in-cloud air at the top of the downdraft. This was consistent with the observations discussed above in which the convergence zone was shown to be well within the cloud.

The model was initialized with the following initial conditions:

- Radar reflectivity: $\text{dBZ} = 55$ dBZ
- Cloud droplet concentration in initial parcel: $\rho_y = 0$ g/m$^3$
- Ice particle concentration in cloud: $\rho_I = 0$ g/m$^3$
- Entrainment parameter: $\alpha = 0.25$
- Initial radius: $r = 500$ m
- Pressure level of simulated boundary level: $P_{BL} = 960$ mb
- Pressure level of ground: $P_{GR} = 1000$ mb
- Cloud droplet concentration in cloudy air: $\rho_{yC} = 1.0$ g/m$^3$

The model soundings were in (pressure, temperature, relative humidity) format:

<table>
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<tr>
<th>p(mb)</th>
<th>$z$(km)</th>
<th>$\theta_e$ (K)</th>
<th>T(K)</th>
<th>RH</th>
<th>W (m/s)</th>
<th>$B_T$ (m/s$^2$)</th>
<th>$B_L$ (m/s$^2$)</th>
<th>$\rho_v$ (g/m$^3$)</th>
</tr>
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<tbody>
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<td>-.069</td>
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<td>.69</td>
<td>0</td>
<td>.000</td>
<td>.000</td>
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</tr>
</tbody>
</table>

Table F.2 Model simulation of cell R using a reflectivity of 55 dBZ. RH is relative humidity, W is the downdraft vertical velocity, $B_T$ is the thermal buoyancy acceleration, $B_L$ is the precipitation loading acceleration, $\rho_v$ is the vapor density. The lifted condensation level from the proximity sounding was 900 mb.

Since the reflectivity in this run was typical of the maximum reflectivity in the cell and not necessarily the typical reflectivity in streamlines in the downdraft (see Table F.1), additional runs with lower reflectivity were conducted. The following run was conducted with reflectivities of 51 dBZ. The two runs represent the range of reflectivities that parcels in the real downdraft experienced.
Table F.3 Model simulation of cell R using reflectivities of 51 dBZ. RH is relative humidity, W is the downdraft vertical velocity, $B_T$ is the thermal buoyancy acceleration, $B_L$ is the precipitation loading acceleration, $\rho_v$ is the vapor density. The lifted condensation level from the proximity sounding was 900 mb.

The model downdrafts accelerate to velocities between -6.5 m/s and -9 m/s. This is within the range of observations, which ranged from -6 m/s using dual Doppler methods (Figure F.17) and -9 m/s using combined dual and triple Doppler methods (Figure F.16). The peak velocities in the model downdrafts are reached at lower levels than is observed. The model contains a crude simulated boundary level that is about 400 m above the surface. Since, as seen in the cross-sections through the real downdraft, the surface divergence begins at about 600 m to 800 m, real parcels decelerate at higher levels than model parcels (see also Figure 5.1).

The thermodynamic qualities of the model downdrafts compare well with the observations at the surface mesonet stations presented above in Section c. The simulations produce surface air that exhibits $\theta_e$ values of about 352 K, very near the observed values. The model downdrafts produce almost no $\theta_e$ change at the surface, consistent with the nearly constant $\theta_e$ that was observed. As was the case with the simulations of cell R, the model downdraft arrives at the ground too warm and too dry when compared to the observations. In the model runs, the surface relative humidity increases to only about .7, whereas almost saturated conditions were observed. Vapor densities were about 19 g/m$^3$, whereas the observed values at station 104 and 24 were 21 g/m$^3$ and 20.5 g/m$^3$. The causes for the discrepancies were discussed along with cell R (in Section f of the cell R analysis) and result from the shallow model boundary level and from surface wetting that is not simulated. Nevertheless, the model parcels arrive with significantly greater moisture than those of the simulations of cell R.
The dominant forcing mechanism appears to be precipitation loading. The precise magnitude of this loading is strongly dependent on the precise precipitation water content. The 55 dBZ run corresponds to about 7 g/m³ of precipitation water content and the 51 dBZ run to about 3 g/m³. The loading forcings are scaled proportionally, .069 m/s² and .036 m/s². It is notable that the upward thermal buoyancy accelerations of .028 m/s² that parcels at the top of the downdraft experience can be balanced by mass loading corresponding to 50 dBZ echoes. Any parcels containing more than about 3 g/m³ precipitation water content would be dragged downwards. This explains why the observed convergence zone is so well correlated with the region containing echoes in excess of this value. This also explains the associated streamline evidence in which parcels transition from horizontal to vertical (downward) motion after crossing the 49 dBZ contour in the cell.

As the model parcels descend, they become colder than the environment due to raindrop evaporation and the positive thermal buoyancy disappears. Some of the rain is consumed in this process and this contributes to the slight lessening of the precipitation loading acceleration. However, the total acceleration grows until boundary layer effects are felt. The 51 dBZ run of the model generates an acceleration value of about .04 m/s² which is almost identical to that observed in the presented streamline (which also contains 51 dBZ echoes).

It must be noted, however, that while precipitation loading is the dominant forcing term at any instant (and in fact, the only downwardly directed term at the top of the downdraft), the evaporation of raindrops plays a crucial role in the dynamics of the downdraft. Without this evaporation, a hypothetical downdraft would warm along a dry adiabat as it descended. While this warm descent would rapidly kill any downdraft originating at the 4 km - 5 km levels of cell R, the situation is less clear with shallow downdrafts like this one. Since the observed environment, Figure 3.B.2, is almost dry adiabatic below 800 mb (about 2 km), the downdraft could preserve its initial upwards buoyancy during descent. This would cut the total acceleration sharply, but if enough precipitation water were present, a downdraft could still be sustained. At least 7 g/m³ would be required to generate total accelerations corresponding to the streamline and convergence observations. The reflectivity evidence indicates (through use of the Geotis (1971) Z-M relationship which gives 55 dBZ = 7 g/m³ in Figure 2.B.12) that this was not the case through most of the downdraft region. The streamline presented in Table F.1 in Section e does not exhibit this much precipitation water content despite the fact that the associated air parcels are clearly
experiencing large downward accelerations. Precipitation water contents approaching 7 g/m³ value were scattered in the downdraft, however, and so it is theoretically possible that downdrafts were forced solely by precipitation loading without the aid of evaporation.

However, if raindrop evaporation were not occurring, the downdraft would have very different thermodynamic characteristics than those were observed. In model simulations in which raindrop evaporation is disallowed, the parcels can be dragged to the surface, but they arrive with a temperature of 305 K and .51 relative humidity. Such parcels are much warmer and drier than the observations, and evaporation of raindrops during surface transit to the mesonet station cannot be reasonably invoked to eliminate the entire discrepancy. As noted before, hot outflows were never observed during the 1986 study period. If hot, dry outflows occurred with any regularity, they would occasionally occur directly over surface mesonet stations and be observed.

Warm outflows were not observed by any of the mesonet stations, implying that strong evaporation was occurring. While it is possible to generate outflows with only precipitation loading, the predicted behavior of such outflows are not consistent with the observations of the outflow of the downdraft in cell F. Raindrop evaporation was necessary to bring consistency between the observations (radar and thermodynamic) and the model results.

Model simulations including cloud water indicate that while the cloud droplets do evaporate preferentially (compared to raindrops), they do not contribute significantly to the downdraft dynamics. The amount of cloud droplets near the base of clouds is observed to be low and much of this shallow downdraft exists below cloud base (at 900 mb, as previously noted). A model run was conducted in which cloud droplets with a concentration of 1 g/m³ were entrained into the downdraft parcels during the time that they are above cloud base (using dBZ = 51 to correspond with the streamline). This run best simulated the observed radar and surface characteristics of the downdraft. As would be expected, the addition of cloud droplet evaporation produced a somewhat cooler and stronger downdraft. Peak downward velocities of nearly -8 m/s correspond best with the best estimates of -8 m/s to 9 m/s in the combined triple Doppler - dual Doppler observations. Surprisingly, the model outflow air, while containing the correct high θ_e, was actually slightly warmer and drier than the outflow air in the runs with no cloud water. This was due to the increased strength of the downdraft which reduced the time available for the evaporation of raindrops below cloud base. Cloud drop evaporation contributed
only 25% of the total evaporational cooling. In runs with reflectivities of 55 dBZ, this proportion was only 17%.

As previously mentioned, ice was not observed at the levels of the storm where the downdraft occurred (the ZDR fields, uniformly high, are not presented), so its role was not explored further.

g. Summary

Multiple Doppler analyses have shown that the downdraft in cell F had its origins at 2 km altitude. Streamlines indicate that air that had experienced a long residence time in the cloud was entrained into the region of the cell containing large reflectivities in excess of 50 dBZ. The parcels descended in this high reflectivity region to the ground. Maximum observed downward velocities were between -6.5 m/s and -9 m/s at about 1 km above the surface.

At the surface, an outflow was observed by radar and several mesonet stations. The thermodynamic characteristics of the outflow were very different from that of a nearby outflow from cell R. While the air from the cell R downdraft exhibited low $\theta_e$ and low $\rho_v$, the air in the downdraft of cell F contained high values of both of these quantities.

Simple model simulations reproduce the strength of the downdraft and the observed accelerations (as measured from horizontal convergence aloft and from directly calculated vertical velocities along streamlines). The model simulations suggest that precipitation loading and evaporation of raindrops are the dominant forcing mechanisms in this downdraft. Cloud droplet evaporation was probably occurring, and model simulations containing cloud droplets produced slightly more realistic downdrafts, but the cooling from this evaporation was small when compared to the cooling from raindrop evaporation. Quite realistic downdrafts and outflows could be generated when cloud droplet evaporation was omitted. Ice phase precipitation was not observed; therefore, cooling due to melting of ice can be ruled out as a forcing mechanism for this downdraft.
3. Cell S

a. Introduction

This Section describes the structure and physics of the downdraft in the cell labelled S in Figure 3.C.1. The downdraft had origins at low levels during the early history of the cell. During the latter phase of the storm, the downdraft deepened and had origins at about 4.5 km. Observations and model simulations will show that both the shallow and deep downdrafts were forced primarily by precipitation loading and the evaporation of rain drops. Differential reflectivity measurements, ZDR, that will be presented later in Section d, show that liquid phase precipitation was predominant in the downdraft region except for evidence of a very narrow mixed phase region at one grid time.

b. Cell S History

The time history of reflectivity in this cell is shown in Figure S.1. The cell develops rapidly after 2013 Z. Maximum reflectivities, as measured by the CP4 radar, quickly exceed 60 dBZ through much of the cloud below 4 km altitude. High reflectivity values, in excess of 50 dBZ, exist above the melting level from 2017 Z onwards. The core of highest reflectivity descends from 3.5 km at 2019 Z to the ground by 2027 Z. The high values of reflectivity in this cell, when compared to the nearby cell F and cell S, imply large precipitation water contents and possibly hail. The Geotis (1971) Z-M formula predicts mass densities in excess of 15 g/m³ in the regions with 60 dBZ or more of reflectivity, but the validity of the formula is questionable at these high values of reflectivity. It is notable that while the precipitation core below the melting level descends, reflectivity continues to rise in the upper reaches of the storm. It appears that there is a decoupling of the downdraft region of the cell, which is confined below 4.5 km at its deepest, from these upper regions.

Figures S.2a and S.2b show horizontal slices through the CP4 measured reflectivity field of cell S, at altitudes of 600 m and 4 km, at three minute intervals during the observed storm history. The proximity of both cell F and cell R, immediately to the north, makes interpretation of the early, low reflectivity, field difficult. Near the surface, at 600 m, the cell is clearly identifiable at 2019 Z from (55,48) to (65,48) with peak reflectivities of about 40-45 dBZ. Tracing backwards towards 2016 Z and 2013 Z, there are hints of earlier development in the bulge of reflectivity near (60,48). The near surface maximum reflectivity increases rapidly, consistent with Figure S.1, to 56 dBZ by 2022 Z and 64 dBZ.
by 2025 Z, heralding the arrival of heavy precipitation at the surface (the phase of which will be discussed later). The near ground level reflectivity drops somewhat after 2028 Z but still remains quite high, maintaining 58 dBZ by 2034 Z.

At 4 km above sea level, cell identification is easier at the earliest times as seen in Figure S.2b. The cell can be seen to rapidly intensify from 30-35 dBZ at 2013 Z to 62 dBZ by 2019 Z. The extremely high values of reflectivity diminish after 2025 Z, although values in excess of 55 dBZ remain. The CP4 radar did not scan through a large portion of the cell at 2031 Z and 2034 Z, so the presented data for those times is as measured by the FL2 radar at 2031 Z and the CP3 radar at 2034 Z. The high reflectivities that are seen by the CP3 radar at 2034 Z indicate that the diminishment of mid level reflectivity plotted in Figure S.1 may be exaggerated.

Surface divergence grows rapidly as the cell develops. Figure S.3 shows surface divergence and near cloud base, 1 km, reflectivity. At 2022 Z, there is almost no surface divergence under the cell. There is a strong region of divergence to the north that is associated with cell R. Mesonet station 27 (see Figure S.5) is under the highest reflectivity region of cell S at this time at about (60,54). At 2025 Z, surface divergence has appeared, reaching peak values of near .011 /s. Mesonet station 112 is under the cell at this time at location (61,60). By 2028 Z, the divergence has strengthened and peak values reach .016 /s. The divergence is located to the rear of the cloud, away from the highest reflectivity region. While the peak effects of the outflow have passed mesonet station 112, now located at (52,60), it is experiencing heavy rain. The outflow continues to strengthen and by 2031 Z exhibits surface horizontal divergence of .02 /s. Mesonet station 33 is at the edge of the outflow at (61,55) and station 32 is under the center of the cell at (57,67). Station 114 is just to the east of the divergence region at (65,60). At the final analyzed time, 2034 Z, the divergence has maintained intensity with peak values of .02 /s. Station 114 is under the peak divergence region at (57,60) at this time.

The history of outflow intensity, as measured by differential radial velocity, $\Delta V$, across it, is shown in Figure S.4. Consistent with the multiple Doppler calculations, there is little diffluence before 2022 Z. $\Delta V$ increases steadily after that time reaching peak values of approximately 24 m/s just after 2034 Z.
c. Surface Mesonet Observations Below Cell S

Figure S.5 shows the Huntsville surface mesonet array, with the positions of the southernmost four radars indicated. The labelled gridpoint locations correspond to the radar grid locations in all figures, normalized to the 2028 Z shifted location. Radar grid locations for other times are written above the plot. (Since the radar grid is shifted eastwards with the cell, the ground and therefore the mesonet array shifts to the west with time.) The location of the surface divergence region, and the smaller regions of maximum divergence, at 2025 Z, 2028 Z, 2031 Z, and 2034 Z, corresponding to the times shown in Figure S.3, is indicated by the enclosed areas. Since there was no measurable surface divergence at 2022 Z, the region of highest reflectivity is indicated instead. As mentioned above, the outflow crossed stations 27, 112, 32, 33, 114, and after the study period, 115, 37, and 38. The records from these stations are discussed below and plotted in Figure S.6.

The collected mesonet time series show that the outflow had two distinct characteristic signatures. During the early stages, the mesonet stations reach saturation while maintaining nearly constant $\theta_e$ and high values of vapor density. At stations that experience the later phase of the cell, the relative humidity, $\theta_e$, and vapor density drop during the height of the event, even when heavy rain is occurring. This differing behavior will be shown to be linked to the evolving downdraft structure aloft. The two distinct phases can be compared with the behavior of cell F (similar to the early cell S outflow) and cell R (similar to the late stage cell S outflow).

Mesonet station 27 experienced the earliest stages of the outflow. As discussed above, and as visible in Figure S.3, there is almost no surface divergence at the time that the cell crosses over this station. In actuality, the outflow develops quickly after the nominal 2022 Z time shown in Figure S.3. The record at station 27 shows that the air passing the station becomes wetter, but not saturated, as the cells passes. Maximum relative humidities reach only .86. The weakness of the outflow is evident in the peak one-minute average winds, which reach only 7-8 m/s at 2023 Z. $\theta_e$ increases slightly during the event and $\rho_v$ rises from less than 18 g/m³ to a moist value of about 21 g/m³. The value of $\theta_e$ during and immediately after the event is about 2 K higher than the 15 minute average taken before onset.

Station 112 was struck about three minutes later at 2025 Z. The record is very similar, though indicative of a slightly more intense event. Relative humidity values reach almost
1.00 and the peak one-minute average winds are over 8 m/s. In a pattern that is very similar to station 27, the values of $\theta_e$ first rise, but then settle at levels near to the long term pre-event mean. $\rho_v$ also rises from less than 18 g/m$^3$ to a moist value of about 21 g/m$^3$. While peak rainfall rates at station 27 were only .36 mm/minute, the values at station 112 peak at about 1.2 mm/minute.

The cell outflow crosses station 32, not plotted, at about 2027 Z. $\theta_e$ remains almost constant during the event while $\rho_v$ rises from 18 g/m$^3$ to about 21 g/m$^3$. Peak rainfall rates are over 2 mm/minute.

Station 33 is the first to exhibit the different behavior that is characteristic of the later stages of the outflow. While the $\theta_e$ and $\rho_v$ histories are similar to that shown by the previous stations during the first minute of the outflow, there is a sharp dip in both quantities starting at 2031 Z. The drop in $\theta_e$, only about 4 K, is not very large when compared to the values immediately before the outflow. It is not as large as the decreases that will be experienced by stations 114, 37, 115, or 38 which are affected later. This is most likely due to the extreme southerly location of station 33 relative to the outflow (see Figure S.5). Air from the downdraft has had to travel along the ground for about 2 km before being measured at this station; during this time it could have become warmer, moister, and diluted with surface air. Surprisingly, the maximum one minute winds were observed to be a relatively high 13 m/s.

Station 114 is strongly affected by the downdraft after 2030 Z. Peak one minute winds are over 10 m/s at four observing times and peaks at 12 m/s. Interpretation of the $\theta_e$ record is not straightforward. Stations 114 and 115 are Lincoln Laboratory mesonet stations. Corrections to the relative humidity record are necessary at these stations as detailed in Appendix 3. Both the uncorrected and two corrected histories of relative humidity, $\rho_v$, and $\theta_e$ are plotted in the figure. Corrections have been applied to all times after 2034 Z. The middle plotted lines which assume that the station gradually dries after the cessation of rainfall, are the most plausible approximations to reality. Just before that time, in the raw record, the relative humidity drops from .95 to .90 and $\rho_v$ and $\theta_e$ drop commensurably. The vapor density has dropped from 21.5 g/m$^3$ to 20.4 g/m$^3$ despite the rain falling at a rate of over 1 mm/minute. Following the middle time series in the figure, the value of $\theta_e$ continues to drop to a low of 345 K at 2040 Z. This value is 8 K lower than the pre-event average. Minimum vapor densities of 19 g/m$^3$ are also evident at 2040 Z. This history
should be compared with that of stations 24 and 104 during the passage of cell R as shown in Figures R.7 and R.8.

The cell passes over stations 37 and 115 after the triple Doppler study period. Station 115, a Lincoln Laboratory site, has been adjusted as per Appendix 3 (corrected and uncorrected values are plotted). The record shows the same characteristic pattern as that observed at station 114. \( \theta_e \) first increases and then drops sharply to values that are 8-10 K below both the pre- and post-event values. The Appendix 3 adjustment accounts for only the last 2 K of this drop. Peak winds are strong, about 12 m/s. The vapor density increases to 21.5 g/m\(^3\) early in the event before dropping quickly to 20.2 g/m\(^3\). The adjustments for persistent saturation only contribute to 0.2 g/m\(^3\) of this drop. Rain is not observed during this event probably due to instrument malfunction. The late stage outflow signature is repeated at the nearby station 37. \( \theta_e \) and \( \rho_v \) drop during 1 mm/minute rainfall rates. \( \theta_e \) reaches values, near 348 K, that are 8-10 K below the pre- and post-event average values.

The record of the event at station 38, while well after the radar study period, shows the most archetypical late stage signature. \( \theta_e \) drops to 346 K, 8-10 K lower than either before or after the event. During intense rainfall, 1.5 mm/minute, the relative humidity drops from .86 to .75 and \( \rho_v \) drops from 20.1 g/m\(^3\) to 18.3 g/m\(^3\). Peak winds are only 8 m/s, indicating that the cell may be weakening or that the station has not been directly hit by the core of the downdraft outflow.

Figure S.7 and S.8 summarize the history of the outflow. Figure S.7 shows the maximum one minute average winds with the maximum values along the station 33-37 axis. Figure S.8 show the contiguous region in which all stations experienced \( \theta_e \) drops of at least 8 K.

d. History of Cell S as Deduced from Triple Doppler Analyses

Radial radar data were loaded into three-dimensional Cartesian grids representing three minute periods. The periods corresponded to average times of 2022 Z, 2025 Z, 2028 Z, 2031 Z, and 2034 Z. Radar coverage was not as complete in the regions affected by cell S as it was in the regions passed by cells R and F. As a result, the calculated wind fields are not as accurate as those in the other cells. Nevertheless, much information about the history of the downdraft in relation to the surface observations just discussed could be
retrieved. Figure S.9 shows an example of the radar coverage at a particularly poorly sampled time, 2025 Z. Only one radar sampled the southeastern portion of the cell, while only three radars sampled in the remainder of the storm with the exception of a small region at the extreme north, which was sampled four times.

In the discussions of radar data that follow, the motion of the cell \((u,v) = (10,3)\) will be subtracted from all wind fields so that the plotted winds are cell relative.

i. 2022 Z.

The downdraft was just beginning at 2022 Z. The high reflectivity rainshaft had just reached the ground (Figure S.1) and the cell was passing over mesonet station 27 (Figure S.6). Surface divergence was very small (Figure S.3).

Horizontal slices through the cell at 1 km and 2 km altitude are shown in Figure S.10 and S.11. The small surface divergence that is present at this time is manifested in the small, -1 m/s at (60,52), downdraft that is calculated at 1 km from the upwards integration of mass continuity from the ground. The vertical winds calculated directly using triple Doppler methods are fairly consistent with these values, near 2 m/s, in the highest reflectivity regions of the cell. Elsewhere, the integrated and directly calculated fields are uncorrelated. This is not surprising due to the inaccuracy of the triple Doppler calculation at this low level. At 2 km, strong convergence is evident through much of the cell, but it is not associated with a downdraft since both methods of vertical velocity calculation indicate upwards air motion at 5 - 10 m/s.

The downdraft is confined to altitudes below 2 km. Figure S.12 is a vertical slice through \(x = 60\) in the previous figures. The strong convergence, up to -.014 /s, extends from near cloud base, at 1 km, to about 3 km. Interpretation of the presented vertical wind fields requires examination of the number of observing radars and the three-dimensional divergence field, both shown in Figure S.13. Five radars looked at the cell at altitudes of 800 m to about 1.8 km. In this region the three-dimensional divergence, a measure of windfield error, is fairly small. Above the highly sampled region the three-dimensional divergence grows in magnitude. This is in contrast to the fields shown in Chapter 2 where this error field decreased with altitude. In the region with low three-dimensional divergence, the vertical winds (unfiltered) are near zero at 1.6 km altitude and positive just above that level. The gridpoint to gridpoint variation is fairly small. In the regions aloft, at
3, 4 and 5 km, the vertical velocities are strongly positive or negative (often overflowing the computation) and exhibit extreme gridpoint to gridpoint noise. The result is that the upwardly integrated vertical velocities, which show a strong updraft above 2 km, are probably the most accurate at all levels except very near 1.6 km where both methods of calculation agree qualitatively. These fields confirm that any weak downdraft that may be present is confined below 1.6 km altitude.

ii. 2025 Z.

By 2025 Z, the downdraft has grown in strength and depth, though it was still not nearly as deep as the downdraft in cell R at this time (see Figure R.13, for example). It was affecting mesonet station 112 as seen in Figure S.6. The shallow nature of the downdraft, as revealed by radar and the nearly constant \( \theta_e \) observed at the affected mesonet station, is similar to that of the downdraft in cell F.

As was noted before (in Figure S.9), radar coverage was not excellent at 2025 Z. A vertical slice through the downdraft, Figure S.14, does reveal the evolution in downdraft structure, however. In this slice, the horizontal convergence, previously observed at low levels from 1 km to 3 km, is strongest from 2.5 km to 4.8 km, and it extends from 1 km to 5 km. The downdraft, best calculated at low levels by using integrated vertical velocities, extends from the ground to only about 2 km. Peak negative velocities are about -3 m/s over a broad region. At higher levels, the integrated fields become unreliable while the directly calculated fields reveal an updraft, extending to high levels in the cloud with a peak strength of over 12 m/s. While only three radars sampled the mid level regions of the cell, their geometry was more favorable than the sampling radars at 2022 Z and thus the directly retrieved winds were reliable. The number of sampling radars and the three-dimensional divergence field, small above 1 km, is shown in the vertical slice in Figure S.15. Since the three-dimensional divergence is small, below 2 m/s/gridpoint or .01/s through much of the region above 2 km, it is inferred that the vertical air motions are correctly calculated in the right panel of Figure S.14.

Differential reflectivity, ZDR, measurements indicate that the bulk of the precipitation water content in the cell below 4 km was in liquid form. Thus, the melting of ice was not affecting the air in the downdraft which, as shown above, did not extend above 2 km. Specifically, Figure S.15 shows the ZDR field in the downdraft region. It is well
correlated with the reflectivity measured by CP2 and is over 3 dB in almost all of the region containing over 55 dBZ of total reflectivity.

iii 2028 Z.

The cell has continued to move east through the study area. No surface mesonet stations were being affected at this time, see Figure S.5, but radar coverage continued. The downdraft now exhibits evidence of a deeper origin level. The eastern portion of the downdraft continues to be shallow, however, and it is this portion which first impacts on each mesonet station. Stations that experience the outflow just a few minutes from this time will be measuring air that is in the downdraft at 2028 Z.

Figure S.16 is a vertical slice through the deepest portion of the downdraft. Both the direct and integrated vertical air velocity fields indicate a downdraft that originates at about 3-4 km and extends to the ground with peak downward velocities of about -8 m/s at 2 km altitude. Other slices through the downdraft show peak downward velocities of about 9 - 10 m/s. Of course the directly calculated fields are not meaningful near the ground and the integrated fields are increasingly unreliable with increasing height, but from 1 km to 3 km altitude they are in quantitative agreement in the downdraft region. The level of the top of the downdraft is most reliably determined by using the directly calculated field. Convergence is occurring from 2 km to 4 km altitude in the high reflectivity region of the cloud and extends to the southern edge of the reflectivity. Figure S.17 shows a slice through this region of convergence at 4 km altitude. The convergence appears to be associated with an engulfment of air into the reflectivity edge at the southern end of the cell near (58,58) (recall that the displayed wind vectors are cell relative winds which have been corrected for the (u,v) = (10,3) motion of the cell). The number of scanning radars is also shown to assure that the convergence zone is not an artifact of combining velocities from regions with four sampling radars with those that have only three or two. The convergence is not oriented along any of the scanning region boundaries. (For an example of a grid that has a bogus convergence zone caused by a sampling boundary, see Figure S.24 at an altitude of 4.5 km.)

The existence or non-existence of ice near the top of the downdraft can be tested by examining the differential reflectivity, ZDR, fields in the downdraft region. Figure S.18 shows both the ZDR and the vertical particle velocity fields. Particle motions are downwards throughout the upper regions of the downdraft. The ZDR field is incomplete.
due to the scanning strategy of the CP2 radar which missed the southeastern portion of the cell. What data there is shows that in the high reflectivity region the ZDR values are above 2 dB, indicating dominantly liquid phase precipitation.

Figure S.19 is a vertical slice about 1 km to the east (or front, relative to the cell motion) of the deepest portion of the downdraft. In this slice the downdraft appears more shallow, extending to only 2.5 km. The peak downward velocities are about -5 m/s in the integrated vertical winds and about -7 m/s in the directly calculated winds which are reliable down to about 1 km due to the favorable geometry and radar coverage present in this region. The major region of convergence extends from 1.6 km to 3 km.

Comparison of the radar data at 2028 Z with the mesonet record at station 33 is illuminating. The wind speed begins to increase at the surface just after 2028 Z, eventually peaking at 13 m/s at 2032 Z. The vapor density first increases sharply while the station is under the shallow portion of the downdraft and then drops sharply for 2-3 minutes while the station is experiencing the outflow under the deepest portions of the downdraft. \( \theta_e \) remains constant during the first minutes of the event and then drops sharply when air from the deepest regions reaches the station at 2031 Z. The early behavior is similar to that of stations that experienced the cell F outflow and the later behavior parallels that of stations under cell R.

iv. 2031 Z.

By 2031 Z, the deepest portion of the downdraft has assumed similar characteristics to the downdraft in cell R. It extends to over 4.5 km altitude and causes significant drops in \( \theta_e \) and vapor density at the surface in its outflow.

Figure S.20 shows a vertical slice through the downdraft at this time showing the usual fields of horizontal divergence, integrated and directly calculated vertical air parcel velocities. Strong convergence is occurring from 2-2.5 km up to and above the level of the top of the downdraft which is near 4.5 - 5.0 km in both the direct and integrated fields. The peak downward velocities are about -8 m/s in a broad region from 1 km to 3.5 km in the integrated fields. In the region of reliability of the directly calculated fields, above 2 km, the maximum velocity is also -8 m/s.
The interpretation of the differential reflectivity field in the downdraft region at this time is difficult. Figure S.21 is a vertical slice showing ZDR, vertical precipitation particle velocities, \( w_p \), and three-dimensional divergence squared, \((V\cdot V)^2\). ZDR values are high through most of the region below 5 km altitude which contains high values of total reflectivity. There is, however, a thin region of lower values from \((y,z) = (69,6)\) to \((y,z) = (69,20)\). The CP2 radar scans the downdraft region with a series of RHI tilts. These tilts are separated by 3 degrees in azimuth so inter-tilt filling of data (described in Chapter 2) is applied in order to calculate the field presented in Figure S.21. The spacing between tilts is \((x,y) = (1.0,2.4)\) km in the downdraft region. The raw, radial format, radar data (Figure S.21b) does suggest that a region of low, near 1 dB, region of ZDR exists at the northern edge of the high reflectivity region in the cell. The depressed values occur in an extremely narrow, less than 1 km, region at 17 km range. This feature shares some similarities with the inferred mixed phase precipitation cores shown in Wakimoto and Bringi (1988).

Difficulties in interpretation arise because:

1. The low ZDR region is not observed at earlier times when the downdraft is already deepening (see Figure S.16). This may indicate that it is a very transient feature. On the other hand the narrowness of the feature compared to the inter-tilt spacing suggests that it may just be missed between tilts. A tilt taken in virtually the same storm relative location (also shown in Figure S.21b) three minutes later does not show the feature, however.

2. The low ZDR region in Figure S.21 is capped at 4-5 km altitude with a region of higher ZDR. If the region really represents a shaft of mixed phase melting precipitation, then what could cause the high ZDR values near the melting layer? It is possible to imagine melting ice phase particles that contain asymmetric regions with liquid water, resulting in non-zero ZDR. But, it is difficult to imagine how, as these particles melted further, they would subsequently exhibit depressed values. The low ZDR shafts discussed in Wakimoto and Bringi (1988) are more contiguous with low ZDR regions above the melting level.

3. The location of the highest reflectivity region is different when measured by the CP2 or the CP4 radars. The downdraft in Figure S.20 is centered on \(y = 69\), the precise location of the low ZDR region. The downdraft is located in the central regions of the cell as measured by CP4. But the low ZDR region is located near the edge of the high reflectivity region as measured by CP2. It is possible that storm evolution has
changed the reflectivity field between the CP2 and CP4 observations. If so, the
evolution has probably changed the velocity structure of the storm during this period,
complicating the interpretation of the gridded Cartesian velocity field which is
constructed with data taken at both times.

4. The horizontal scale of the low ZDR region is smaller than the horizontal scale of the
downdraft. In Figure S.21, the region in which precipitation is moving downwards
at nearly 10 m/s has a horizontal scale of at least 2 km. The ZDR depression spans
less than 1 km in Figure S.21b. However, the velocity feature may have been spatially
smoothed since the velocities are calculated from data taken at many different times.

5. Horizontal convergence is not well correlated with the low ZDR region. In
Figure S.20, the maximum horizontal convergence appears to be associated with air
parcels that are entering the highest reflectivity region from the south (left). The
maximum convergence is to the south of the low ZDR region.

6. The low ZDR region is not associated with enhanced total reflectivity. If the
precipitation particles in the low ZDR region were comprised of melting ice particles
the increased dielectric constant of the partially melted particles would probably cause
some increases in total reflectivity.

Despite these problems of interpretation, it is difficult to deny the raw radial data which
shows the low ZDR region. The region is co-located with radial velocity perturbations
( not shown ) which are related to the triple Doppler calculated downdraft. The
consistency with the observations of Wakimoto and Bringi (1988) and the location,
coincident with the downdraft, lend credence to the feature's veracity.

The more shallow frontal portion of the downdraft has undergone little evolution and is
shown in Figure S.22 to extend to about 2 km in the integrated vertical velocity field. The
directly calculated field confirms that air above 3 km is moving upwards and, at the lowest
limits of its reliability, just begins to show the downdraft. Horizontal convergence is
occurring throughout a deep layer of the cell, extending down to 1 km altitude.

v. 2034 Z.

At 2034 Z, the last triple Doppler observation time, the downdraft continues to be both
strong and deep. At the surface, mesonet station 115 is about to be affected and will
experience a sharp $\theta_e$ drop.
Figures S.23 and S.24 show the final vertical slices through the deepest portion of the downdraft. The slices extend from the ground to 6 km, but triple Doppler and reliable dual Doppler coverage do not extend above 4-5 km altitude as seen in Figure S.23. The extreme values of horizontal divergence at this level, Figure S.24, delineates the upper limit of reasonable data. If the abrupt topping off of the vertically integrated vertical velocities is ignored, then the top of the downdraft can be seen to be above the level of radar coverage. The triple Doppler show that the downdraft extends to 5 km altitude. As shown in Figure S.24, the peak downward velocities in the integrated vertical air velocity field are near -10 m/s. This is significantly different that the directly calculated peak vertical air velocity, -6 m/s. The mid level convergence that was visible at 2028 Z and 2031 Z is not evident. Perhaps it has disappeared or perhaps it has just not been sampled by the poorly scanning radars.

Precipitation motion is strongly downward throughout the sampled region of the downdraft shown in Figure S.23. Differential reflectivity measurements, also shown in Figure S.23, indicate that the dominant precipitation phase is liquid. ZDR values below 1 dB rarely penetrate the 40 dBZ contour. The narrow region of low ZDR that was observed at 2031 Z is absent at this time. The region with values of ZDR near zero extends as low as 5 km in the cell. While this may result in ice falling into the downdraft, this is impossible to confirm since, in the region of lowest ice penetration near 5.2 km, (70,26) , the nearest region of calculated vertical velocities is 600 - 800 below at 4.6 km, ( 70,23). Air that may be cooled by possible ice near the top of the downdraft will not reach the surface until at least 2044 Z at the observed vertical air velocities.

The frontmost region of the downdraft continues to be about 2.5 km deep and contain moderate downward velocities. A vertical slice to the east of the deep downdraft is shown in Figure S.25. The convergence at 1-3 km altitude is still present. The integrated vertical wind field suggests a downdraft originating at 2.5 km, while the direct calculation confirms that air at 3-4 km altitude is moving upwards, providing an upper bound to the height of the downdraft in this region.
e. Streamlines

Streamlines were calculated and examined in order to measure accelerations following air parcels. A typical streamline tracing air entering the downdraft is shown below in Table S.1.

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Table S.1 Streamline entering downdraft of cell S at 2031 Z. Various properties of an air parcel are listed. Time in seconds, \((\text{x,y})\) locations in radar grid locations (1 gridpoint = 200 m), \(z\) in km, \((u,v,w_a)\) velocity (m/s) reflectivity from the CP4 radar, dBZ, precipitation water content (g/m³), \(M\), changes in precipitation water content, dM, and changes in vertical velocity, dw_a.

Because of the poor quality of the vertical velocity fields, trajectories could not be calculated. Instead, short streamlines, which implicitly assume that all physical fields remain constant in time, were traced. In the presented streamline, the parcel accelerates downward as the reflectivity and precipitation water content increases. The parcel accelerations are approximately -.05 - .10 m/s², about twice as large as the accelerations observed in cell R and cell F. Precipitation water contents exceed 4.5 g/m³.

f. Forcing Mechanisms of the Downdraft in Cell S

Simple calculations of the magnitude of the various forcing mechanisms in cell F are shown in the first Section below (i). Discussion of the results of a one-dimensional parcel model are presented in the next Section (ii). The one-dimensional model is more thoroughly discussed in Appendix 2.
i. Calculation of the Instantaneous Forcing of the Downdraft in Cell F

Following the discussion of cell R, the acceleration caused by the cooling of the
subsaturated engulfed air from the south of the cell could cause negative buoyancies over
.14 m/s² if the engulfed air is given over 10 minutes to saturate. This is much longer than
the observed travel time into the downdraft. Realistically, accelerations due to negative
buoyancy of the order of .05 m/s² could occur, as will be seen in the model simulations
presented below. This is not as large as the observed accelerations and thus this
mechanism alone could not be responsible for the origination of the downdraft motion or
the rapid increases in the acceleration rate that are observed in streamlines entering the top
of the downdraft.

Accelerations caused by precipitation loading are large, near .08 m/s². These are large
enough by themselves to account for the observations in the streamlines.

Estimation of cooling due to melting of the ice that may be present in the narrow low ZDR
region in Figures S.21 and S.21b is more difficult. In Chapter 1, a simple calculation
produced coolings of 5 K from melting ice. The observations suggest that the amount of
ice phase precipitation is less than in that calculation and that the region of air that must be
cooled is much larger than the region containing ice.

Through much of the low ZDR region, radar reflectivities are near 50 dBZ. The
Geotis (1971) Z-M relation suggests water contents of 3 g/m³ at this level of reflectivity.
The relation does not apply to ice but, since it is likely to overestimate liquid water contents
in a region of melting precipitation, it provides a useful upper bound to the concentration of
ice. This upper bound should be modified to account for the likelihood that the low ZDR
region contains both liquid and ice phase precipitation. As discussed earlier in the analysis
of cell R, this estimate of ice water content can be checked for consistency with the liquid
water content below the ice or mixed phase region.

Based on a ice water content of 3 g/m³, and following the discussion in Chapter 1, the
melting of ice could cool the low ZDR region by about 2 K. This could produce negative
buoyancy accelerations of .07 m/s², large enough to explain the observed accelerations. As
noted before, the downdraft is at least twice the diameter of the narrow low ZDR region
and the cooling and acceleration should therefore be reduced by roughly a factor of four.
ii. Model Results

The one-dimensional model was run in order to simulate the deep phase cell S downdraft. The model results closely follow those of cell R and cell F and the more detailed discussions of the results in those Sections should be read before the results presented here. The model was initialized using unsaturated air at the top of the downdraft. This was consistent with the observation that air was apparently being engulfed into the cell from the south at about the 4 km level.

Radar reflectivity
Cloud droplet concentration in initial parcel
Ice particle concentration in cloud
Entrainment parameter
Initial radius
Pressure level of simulated boundary level
Pressure level of ground
Cloud droplet concentration in cloudy air

The model soundings were in (pressure, temperature, relative humidity) format:

outside environment (600,277,.55) (800,289,.90) (1000,304,.65)
outside entrainment environment (600,277,.55) (900,294,.75) (1000,298,.95)
cloudy entrainment environment (600,278,1.) (900,294,1.) (1000,298,.95)

Table S.2 Model simulation of cell S using a reflectivity of 53 dBZ and dry air entrainment. RH is relative humidity, W is the downdraft vertical velocity, $B_T$ is the thermal buoyancy acceleration, $B_L$ is the precipitation loading acceleration, $\rho_v$ is the vapor density, $Z$ is the altitude in km.
Table S.3  Model simulation of cell S using a reflectivity of 53 dBZ and cloudy air entrainment. RH is relative humidity, $W$ is the downdraft vertical velocity, $B_T$ is the thermal buoyancy acceleration, $B_L$ is the precipitation loading acceleration, $\rho_v$ is the vapor density.

The shallow early phase of the downdraft as well as the frontmost portion of the late stage downdraft existed in an environment very similar to the cell F downdraft. The model simulation of that downdraft, Table F.2, applies here as well and is not reprinted.

The downdrafts in the simulations of the deep phase of cell S are more intense than those in the simulations of cell R. This is due to greater precipitation mass loading. The simulated vertical velocities compare well with the observations. The peak model and observed velocities in the deep downdraft are about -8 m/s. The model produces strong velocities just above its thin boundary level that are greater than the observations. Possible reasons for this are discussed along with the model simulations of the cell R downdraft earlier in this chapter. The simulation of the shallow portion of the downdraft (Table F.2) produces a peak downward velocity of -9 m/s just above the boundary level and -6 m/s near cloud base. The observations show a -5 m/s to -6 m/s peak velocity near cloud base.

The thermodynamic qualities of the model downdrafts compare well with the observations. The shallow downdraft simulation produces surface air with a high $\theta_e$, 352 K, and a high vapor density that are consistent with the observations at stations 112 and 27 which were affected early in the downdraft history. The deep downdraft simulations both produce surface air that has low $\theta_e$ (349 K dry entrainment, 340 K cloudy entrainment) and low vapor density (14.5 g/m$^3$ dry entrainment, 18.2 g/m$^3$ cloudy entrainment). These low values are consistent with the observations at the stations hit by the late stage outflow. The dry entrainment run produces values of $\theta_e$ and vapor density that are as low as that observed by any mesonet station while the cloudy entrainment run fails to account for the minimum $\theta_e$ observations at stations 38, 342 K, and 114, 346 K. Streamlines and the broad convergence zone indicate that entrained air had a variety of origins and that some
mix of the entrained environments in the two simulations is the closest approximation to reality.

The role of melting ice in the narrow low ZDR region was explored with the model. The model was initialized with 50% ice water content. The melting rate in the model was set at a level which produced a deep mixed phase layer. The model did not accurately simulate the full depth of the observed mixed phase region as it tended to produce a more limited melting layer.

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<th>T (K)</th>
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Table S.4 Model simulation of cell S using a reflectivity of 53 dBZ and cloudy air entrainment and 50% ice water content. RH is relative humidity, $W$ is the downdraft vertical velocity, $B_T$ is the thermal buoyancy acceleration, $B_L$ is the precipitation loading acceleration, $\rho_v$ is the vapor density.

The ice has a secondary effect on downdraft intensity. When this simulation is compared with that in Table S.3, the minor effects on thermal buoyancy and downdraft velocity can be seen. The melting ice caused 2.5 K of the total cooling experienced by the simulated parcels. This was partially offset by 1.1 K of warming caused by condensation onto the melting particles when the local wet bulb temperature was above 273 K. The model parcels arrive at the ground with values of $\theta_e$ and water vapor density that are very similar to the simulated parcels in the no-ice run.

Mesonet stations that experienced the effects of this cell at the later study times experienced both the shallow and deep portions of the late stage downdraft. Thus, station 33 exhibited high values of $\theta_e$ and increasing vapor density during the early portion of the event, consistent with the model run presented in Table F.2. Later, it exhibited a drop in both quantities, consistent with the deep downdraft simulations presented in Tables S.2 and S.3. This basic pattern was repeated at several of the plotted mesonet station records.

The dominant forcing mechanism in both the deep and shallow simulations appears to be precipitation loading since this is always larger than the instantaneous thermal buoyancy.
but, following the discussions of both cell R and cell F, this is misleading. In the absence of raindrop evaporation, parcels would become unrealistically warm as they descend and would become extremely upwardly buoyant. Even with the evaporation that is present, the simulated parcels are upwardly buoyant through most of the descent to the ground. As was the case in cell R, to the extent that warm, cloudy air was entrained, cloud drops evaporated and offset the effects of the inclusion of the positively buoyant cloudy air. In the cloudy air entrainment run, about 40\% of the total evaporative cooling was caused by the evaporation of cloud drops. The ice that may or may not be present during much of the deep downdraft's history appears to play only a minor forcing role.

g. Summary

Multiple Doppler analyses have shown that a shallow downdraft forms in cell S. The origin level is at about 2 km at 2025 Z. The downdraft deepens so that by 2031 Z, it has its origins near 4.5 km. Streamline analyses indicate that air is being entrained from the southern side of the storm and accelerating downwards in the precipitation core. Accelerations calculated along streamlines are consistent with the radar inferred forcing that is present. The observed downdraft reaches the ground and has peak downward velocities of about 8-10 m/s.

At the surface, the shallow, early phase of the downdraft is observed by two mesonet stations which both exhibit nearly constant values of $\theta_v$ and high $\rho_v$ during the event, in a fashion similar to stations that experience the outflow under cell F. Stations that experience the outflow at later times exhibit a rapid increase in vapor density with constant $\theta_v$ as the frontmost, shallow portion of the downdraft passes. They experience a decrease in both quantities as the deep downdraft passes overhead. The decreases follow a pattern very similar to those under the downdraft in cell R.

Simple model simulations of both the shallow and deep downdrafts reproduced both the -8 m/s to -10 m/s vertical winds and the varied surface thermodynamic conditions. The model simulations suggest that precipitation loading and evaporation of raindrops are the dominant forcing mechanisms in this downdraft during all phases of its evolution. The evaporation of entrained cloud droplets may have played a secondary role in further cooling warm cloudy air that was mixed into the descending parcels.
The role of melting ice appears to be minor in this downdraft. The single observation of a low ZDR region in the downdraft at 2031 Z, is intriguing in light of similar observations by Wakimoto and Bringi (1988), but both the level of inferred ice water content and its narrow extent compared to the region containing downward moving air prevent it from playing a major role in downdraft forcing.
CHAPTER 4  Orlando Outflow

A. Introduction

A small convective system moved into the radar array in Orlando, Florida see Figure 2.B.5, on 03 October 1990. It produced an outflow at the northern edge of the array. In this Chapter, the downdraft causing this outflow is described and the forcing mechanisms that forced the downdraft are diagnosed. The downdraft is described using multiple Doppler analyses aloft and at the surface. Surface mesonet thermodynamic observations are used along with simple model calculations to diagnose individual forcing mechanisms.

B. Synoptic Setting

The large scale environment in which the storm formed was fairly typical for fall in Florida. The surface and upper level flows were weak. The surface flow was from the east and the 500 mb flow was light and variable. Figure 4.B.1 shows the surface weather at 1800 Z, (about two hours before the downdraft event) and 500 mb environment at 1200 Z (about eight hours before the downdrafts occurred). Temperatures, dew points and $\theta_e$ values were unremarkable. Unfortunately, special proximity balloon soundings taken earlier during the summer oriented experiment had ceased. The nearest soundings taken on the study day were from Tampa, FL (TBW), on the Gulf of Mexico coast and West Palm Beach, FL (PBI) on the Atlantic coast. The soundings at these stations were taken at the standard times, 0000 Z and 1200 Z and were thus well separated from the event time of roughly 2000 Z. The soundings taken at each of the stations at the times nearest the event are shown in Figure 4.B.2. At 1200 Z, in the early morning before the event, the soundings exhibit shallow surface inversions with high values of $\theta_e$ above them. $\theta_e$ is about 350 K at 1000 mb and drops to minimums of about 320 K to 330 K in both soundings. There is a pronounced drying above 650 mb in Tampa and 580 mb in West Palm Beach. Generally, during days that had dry air aloft, there was very little deep convection. The soundings taken at 0000 Z, after the downdraft event, showed that the atmosphere had become saturated between 950 mb and the still present dry layer above about 600 mb. Mid-level temperatures had remained nearly constant, so the increased humidity resulted in a $\theta_e$ increase of about 5 - 10 K. Lifted condensation levels of undilute surface air, LCL, were at about 950 mb or 500 m above the ground. (The LCL at 1952 Z at Orlando airport, MCO, was 910 mb or about 1000 m above the ground.) Both before
and after the event the winds below about 600 mb were from the east at about 5 - 10 knots (3 - 5 m/s), roughly the direction of cell movement, as will be seen in the next Section. Winds were light and variable from 500 mb to 600 mb. Above 500 mb, north to northwesterly winds were evident at 1200 Z and northeasterly winds at 0000 Z.

Convection was widespread across Florida during the afternoon. Figure 4.B.3 is the national radar summary valid at 1935 Z, about 50 minutes before the event. While this summary shows a cell west of Orlando that has a top of 42,000 feet (12.8 km), the radar evidence presented below will show that the cell that produced the studied outflow was confined to levels below 6.5 km (440 mb) except for a very weak echo region, exhibiting reflectivities of less than 15 dBZ, that extended to about 10 km (33,000 feet) (270 mb) (see Figure 4.C.1 at 200735 Z in the next Section).

C. Storm History

1. Reflectivity history
   
a. Early History Using Radial Data

At about 2000 Z a convective system develops to the northeast of the study area. Figure 4.C.1 shows ungridded radial radar data taken prior to the period of triple Doppler study. The scans at 195938 Z and 200551 Z show that a storm or small collection of storms is moving slowly east towards the observing radar (MIT). Between these times a cell develops to the south of the system that has surface reflectivities of 40-45 dBZ. A RHI through this cell, panel c, shows that it has already developed considerable reflectivity aloft, exceeding 55 dBZ at 4 km. A very weak echo extends to 10 km, but the main body of the storm is confined to levels below 6.5 km. Observations taken by the FL2 radar just after the time of that RHI show that the cell had a distinct radar reflectivity signature at the ground.


FL2 measured reflectivities of 55-60 dBZ at ground level, several dB higher than those measured by MIT. Except for this ground level PPI, all other presented observations will use measurements from the MIT radar. None of the three radars had a noticeable systematic bias in their reflectivity measurements, as compared to the other two.
comparisons of many identical grid volumes, MIT measured the median value more often than either FL2 or UND. Thus it was judged the most representative. More importantly, MIT scanned the storm with a tightly spaced PPI pattern with only small inter-tilt gaps. Figure 4.C.2 shows a vertical slice through the gridded reflectivity fields extending from the ground to 6 km through the cell at a later time (2024 Z). The displayed fields are uninterpolated and clearly illustrate the effects of inter-tilt spacing. While the coverage below 1 km is almost complete (The occasional gaps in the MIT grids are due its 250 m gate spacing compared to the 200 m grid spacing. This was somewhat greater than the other two radars, but still very small compared to the inter-tilt gaps and was not the determining factor in determining three-dimensional data resolution.), there are large gaps in the fields measured by FL2 and UND. The spacing in the FL2 data is about 1.2 km and the spacing from UND is about 1.0 km. Comparison with the RHI in Figure 4.C.1 shows that certain cell structures, notably the reflectivity maxima at 1.8 km and 4 km, are comparable or smaller than this spacing. The MIT inter-tilt gaps average less than 400 m throughout the storm volume and would resolve even these small features. The FL2 and UND scanning strategy would not. By using the MIT radar for reflectivity analyses, measured, almost non-interpolated, data can be used through most of storm volume.

c. Storm History During Coordinated Scanning Period Using Gridded Data

Coordinated scanning by all three radars began just after 2012 Z. Figure 4.C.1 shows the ground level scan from MIT, panel e, at the beginning of this period showing that the southern cell had become dominant. (This radial data can be compared to the gridded data shown in figure 4.C.5 for the 2012 Z grid time. The agreement is excellent and most features visible in the raw radial data are visible in the gridded field.) To account for storm motion, the displayed grids have been shifted at a velocity of \((u,v) = (-3,0)\) m/s, as described in Chapter 2. Because of packed data storage limitations (the details of which are unimportant, but, briefly the values of delta-time from the time to which storm motions are reduced cannot exceed +/-128 units where the unit has been set to 10 seconds.), the grids at 2012 Z and 2016 Z were shifted to locations valid at 2020 Z, while the grids after that time were shifted to 2030 Z locations. The net effect is that the former grids are displaced to the east by:

\[
3 \text{ m/s} \times (2030 \text{ Z} - 2020 \text{ Z}) = 3 \text{ m/s} \times 600 \text{ s} = 1.8 \text{ km} = 9 \text{ gridpoints}
\]

compared to the latter grids. Figures in which data from different times are presented display plot areas that have been adjusted to account for this 9 gridpoint shift. Thus the same cell-relative region is always shown for consistency.
Figure 4.C.3 shows the time history of the vertical development of maximum reflectivity in the cell, as computed from the gridded data after 2012 Z. The descent of the reflectivity field is the most striking feature. The upper 55 dBZ contour, at 5.0 km at 2012 Z, descends at 4 m/s to 2.5 km at 2024 Z and to the ground by 2032 Z; the 40 dBZ contour drops at 2.6 m/s from 5.8 km to 1.4 km from 2012 Z until 2040 Z. Peak reflectivities of near 60 dBZ were observed for a brief time at about 4 km, just below the environmental melting level of 4.5 km. The highest reflectivities to reach the ground do so at 2028 Z, which closely corresponds to the time of maximum outflow strength, shown with arrows in the figure. This collapse is also very apparent in a series of vertical slices taken through the same cell relative volume during the 2012 Z - 2040 Z period, Figure 4.C.4. The descent of the high reflectivity core from 4 km at 2012 Z to 1 km at 2024 Z is evident. By 2036 Z, there are only isolated gridpoints exhibiting even 40 dBZ.

A small cell develops to the southwest of the main cell towards the end of the study period. Its reflectivity history is also shown in Figure 4.C.3. The cell is visible in Figures 4.C.5 and 4.C.6, discussed below, as an area of high reflectivity in the (40,80) to (50,80) area in the 2032 Z, 2036 Z, and 2040 Z fields. The cell is short lived, as can be seen by its rapid collapse, beginning at 2036 Z. It produces a weak outflow and by 2046 Z, radial reflectivity data, not shown, indicated that peak surface reflectivities had dropped to about 40 dBZ.

A time series of horizontal slices through the cells are shown in Figures 4.C.5 and 4.C.6. Figure 4.C.5 shows near-ground level reflectivity, which exceeds 54 dBZ from 2020 Z until 2028 Z, during the period of strongest outflow period. Very high reflectivities, up to 60 dBZ, associated with the southwestern cell can be seen at 2036 Z and 2040 Z, to the south of the remnants of the primary storm. Figure 4.C.6 shows horizontal slices at 4 km, just below the melting level, through the same regions at the same times. The rapid collapse of mid-level reflectivity can be seen from 2016 Z to 2024 Z. By 2024, the peak reflectivity value in the primary cell is only 31 dBZ. There is a transient burst of reflectivity, peaking at 2036 Z (46 dBZ), which is associated with the southwestern cell.

2. Surface Outflow History as Deduced from Radar Measurements

Figure 4.C.7 shows surface horizontal divergence (unsmoothed) at each gridded analysis time from 2012 Z until 2040 Z. It is calculated from the overdetermined dual Doppler wind fields. The divergence grows rapidly between 2012 Z and 2024 Z. It is well correlated
with the highest reflectivity region throughout the period. Peak values of up to 0.015/s are common at 2024 Z and 2028 Z. As the reflectivity field in the storm subsides, the divergence weakens. At 2032 Z, the only substantial divergence is under the small high reflectivity region at the east edge of the storm and by 2036 Z and 2040 Z, there is almost no surface divergence at all. The peak surface divergence coincides with the descent of the peak reflectivity values to the ground as shown above in Figure 4.C.3. Figure 4.C.8 shows a time history of the peak surface horizontal divergence (smoothed) and time history of differential radial velocity, ΔV, measured across the outflow. Peak ΔV values of 15 m/s at 2024 Z and 2028 Z correspond well with the divergence calculations. The measurements were made from ground level data taken by the FL2 radar at approximately one minute intervals.

There is some evidence for weak divergence under the high reflectivity region associated with the late stage southern cell that is visible in Figure 4.C.7 at 2036 Z and 2040 Z, but this feature is not of concern in this study.

The region of divergence under the main cell appears to separate into two areas after 2028 Z in Figure 4.C.7. Each is associated with a region of high reflectivity. If these are treated as separate cells then the collapse of the northwestern region is apparently even faster than indicated and the growth and decay of the southeastern divergence region is well correlated with the growth and decay of reflectivity in that cell. Raw radial velocity data from the three radars fail to show distinct surface outflows and these hypothetical cells are difficult to distinguish as separate entities until the final collapsing stages. Therefore the cells and outflows will be treated as portions of a whole in these discussions.

3. Surface Mesonet Observations Below the Orlando Outflow

Figure 4.C.9 shows the Orlando surface mesonet array, with the position of the three radars shown. The labelled gridpoint locations correspond to the radar grid locations in all figures, normalized to the 2030 Z motion reduced location. Since the radar grid is shifted westward with the storms, the ground and therefore the mesonet array shifts to the east with time. The location of the maximum surface divergence, as calculated from radar data, at times corresponding to those in figure 4.C.7, is indicated. The edge of the divergence region can be seen to impinge on stations 22 and 23 just after 2016 Z. This timing is consistent with the observations taken at these two locations, shown in Figures 4.C.10.
and 4.C.11. Station 36 is hit directly by the central portion of the outflow, but, unfortunately, the station is not functioning.

Station 23 feels the strongest effects from the outflow and the record from that station is shown in Figure 4.C.10. The time series extends from 2005 Z until 2030 Z. At the beginning of the period, the station exhibits high $\theta_e$ values near 355 K, a temperature of 302 K and a relative humidity of .80. Just before the outflow, the parcel has cooled and dried slightly so that $\theta_e$ is about 352 K. When the outflow hits, at about 2015 Z, the temperature drops slowly to 298 K, the relative humidity rises to nearly saturation, and $\theta_e$ first rises to 356 K and then falls to about 353 K. (The $\theta_e$ jump is probably caused by different response times for the temperature and relative humidity sensors and is discussed in detail in Appendix 1.) Peak one minute averaged winds of 7 m/s are seen at 2020 Z. This is still four to eight minutes before the time of peak outflow strength measured by the radar.

Station 22 exhibits very similar behavior. Values of $\theta_e$, temperature, and relative humidity, shown in Figure 4.C.11, follow the pattern of station 23. Since, as can be seen in Figure 4.C.9, this station is only glancingly affected by the outflow, the cooling and saturation process is somewhat slower and the winds are weaker. While the plotted one-minute average winds are relatively constant with time, the event is marked by increased peak gust winds (indicated with a dashed line over the one minute winds). The long period of slightly enhanced winds after the event is probably due to the continued proximity of the trailing, and then separate, region of divergence in the southeastern portion of the cell.

The Orlando airport, located within the triple Doppler triangle and indicated in Figure 4.C.9 with parallel runway lines, took hourly surface observations. The observation taken at 1952 Z was $T = 86^\circ$F, $T_{dewpoint} = 72^\circ$F, $P = 1020$ mb. The resultant relative humidity is .62 and $\theta_e = 350$ K. Both these values are somewhat lower than the mesonet readings.

The mesonet observations, particularly the nearly constant values of $\theta_e$, will have to be explained by examining the radar evidence in Section D, and the modelling results in Section F. Unfortunately, the outflow did not directly hit the two stations and it passed them early in the outflow history, at least four minutes before the 2024 Z - 2028 Z peak
strength period. Therefore the constraints placed on the model results are not as strong as those that could be applied with the Huntsville outflows.

D. History of the Orlando Downdraft as Deduced from Triple and Multiple Doppler Analyses

The history of the downdraft is studied with the aid of triple Doppler analyses. The downdraft is seen to be shallow, originating at the 2 km to 3 km level. Air is seen to enter the downdraft producing cell from the north, pass into the high reflectivity region of the cell, and then accelerate downwards in the downdraft to the ground.

Radar data are loaded into Cartesian grids each representing four minute duration snapshots. The nominal time of the grids corresponds closely with the start time of coordinated volume scans conducted by the three radars. Each radar scans the study region with a series of sector PPI’s with varying inter-tilt spacings (as discussed in Section C above). As mentioned before, the grid is shifted at (u,v) = (-3,0) m/s in order to keep the convective system roughly stationary relative to the grid.

Strong horizontal convergence at 4 km altitude is observed from 2012 Z until 2016 Z, at 3 km from 2012 Z until 2020 Z and at 2 km from 2012 Z until 2024 Z. Horizontal cross-sections through the cell at 3 km, Figure 4.D.1, show the horizontal convergence fields from 2012 Z until 2032 Z. (In these fields, data have been eliminated from regions sampled by only two radars.) The convergence exists almost exclusively within the high reflectivity regions of the cloud and disappears when the high reflectivity descends below the 3 km slice level at 2028 Z. The divergence evident to the south of the convergence at (70,80) at 2012 Z is associated with the cell updraft. At 4 km, Figure 4.D.2, the convergence - divergence couplet is clearly visible at 2012 Z and 2016 Z but the field quickly weakens after 2020 Z. The highest reflectivities, over 55 dBZ, are in the divergent updraft. At 2 km, Figure 4.D.3, there is strong convergence, -.011 /s, in the highest reflectivity region at 2012 Z that gradually weakens and disappears by 2028 Z. The association between horizontal convergence and high reflectivity is shown clearly in Figure 4.D.4 in which the vertical extent of the convergence zone is plotted along with the vertical extent of the 55 dBZ contour. The upper boundary of the convergence zone drops with the 55 dBZ contour until after 2028 Z when it follows the 55 dBZ contour in the southwestern cell.
This history suggests a downdraft, with origins at low to mid levels, 2 km - 3 km, in the cell, that becomes more shallow with time as the storm collapses. Convergence alone, however, does not determine the origin level of the downdraft since the air may, and at early times does, converge and rise into the updraft. Triple Doppler and multiple Doppler analyses of the vertical velocity field are necessary to clarify this structure. Analyses for five specific time intervals (2012 Z, 2016 Z, 2020 Z, 2028 Z, 2032 Z) are discussed in turn.

1. 2012 Z

At 2012 Z, near the time of the apogee of the contours of maximum reflectivity in Figure 4.C.3, the cell contains only a small downdraft. While the downdraft is to grow in strength during the next several minutes, it exhibits downward velocities of only about -3 m/s. Figure 4.D.4b shows a north-south vertical slice through the center of the cell showing the vigorous updraft and nascent downdraft and outflow. (The techniques and problems of accuracy associated with the plotted wind fields in this figure and the figures valid at different times during the storm are discussed in detail in the next Section concerning the downdraft at 2016 Z.) The figure shows the horizontal divergence and slice parallel, cell relative, winds as well as vertical winds calculated using triple Doppler and dual Doppler methods. There is a prominent region of convergence from 1 km to 4 km altitude. Air can be seen entering this region and then being accelerated upwards into the updraft (as confirmed by tracing air parcel streamlines). Reflectivities in this region are typically in the 45 dBZ to 50 dBZ range. There is no convergence in the region of highest reflectivity near 4 km. Air destined for the downdraft enters the cell from the north (right) below 2 km and is weakly accelerated towards the ground. At the ground there is a weak outflow and divergence (see also Figure 4.C.7). By 2016 Z and 2020 Z, as the mid-level reflectivity core drops to the ground, the downdraft and the outflow become more prominent.

2. 2016 Z

At 2016 Z, the cell is just past peak vertical development, as measured by the vertical extent of reflectivity. The reflectivity core has just begun to collapse and, as noted in Figure 4.D.4b (which shows the cell at 2012 Z in the previous Section), significant convergence is occurring across a wide range of altitudes. Direct triple Doppler analysis is difficult in this cell because, as can be seen in Figure 4.C.9, the storm is on the periphery.
of the triangle defined by the three radars. As discussed in Chapter 2, low variances in derived vertical winds are found only in the interior of this triple Doppler triangle. In addition, the coarse scanning patterns of the FL2 and UND radars, noted in the previous Section, cause a reliance on interpolated radial data in regions over 1 km in extent. In a small rapidly evolving storm such as this, these gaps significantly affect data quality. Significant manipulation is necessary to produce reliable vertical wind fields. In addition, the grid update rate of four minutes is slow compared to the rapid evolutionary changes occurring in the cell.

To illustrate the data manipulation that took place, several fields will be shown. Figures 4.D.5-4.D.7 all show 4 km by 4 km slices through the cell at 3 km altitude. In Figure 4.D.5, the number of radars that sample each grid point is shown. The displayed grid has been interpolated and so inter-tilt gaps have been almost completely filled. Since coordinated scanning is occurring, the sampling time and sampling volume variation common in Huntsville (see Figure 2.C.2) is not observed. The radars sample this horizontal slice at slightly different times, and the velocities from the respective radars become noisy and unreliable near the cloud edge at slightly different locations. Thus, there are regions in which only two radars record data at the cloud edge. Wind data from these regions are excluded in future figures. Also shown in Figure 4.D.5 is the smoothed (eliminating features with wavelengths of less than 400 m), squared three-dimensional divergence. It is quite large near the edge of the storm, as expected. There is also a region of high values in the convergence region (see figure 4.D.1) near (70,90). As discussed in Chapter 2, large three-dimensional divergence is an indication of inaccurately calculated wind fields. Regions with three-dimensional divergence larger than 4 m/s/gridpoint are excluded in future figures. In addition, because data quality is low near the edge of the cloud, and because the convergence and downdraft are confined to interior regions of the cell, data in regions where the MIT radar measured a reflectivity level of less than 20 dBZ are excluded in future figures. The effect of all this filtering is illustrated in Figure 4.D.6 where the raw, triple Doppler vertical particle velocity field (already effectively filtered to have only values where three radars sample, of course) and the filtered air parcel velocity field are shown. The gaps in the filtered fields can be filled and the data smoothed; Figure 4.D.7 shows the final result for the vertical and horizontal wind fields. The horizontal wind vectors clearly show the horizontal convergence zone. The convergence is well within the cell and the air parcels travel at least 1 km from the reflectivity edge before entering it from the north or south.
While the slices at 3 km altitude are illustrative of the processing conducted on the data from the 2016 Z and other grids, the downdraft is largely confined to lower levels, below 2 km. Figure 4.D.8, a horizontal slice at 2 km altitude, shows the horizontal wind field and the strong convergence within the high reflectivity region of the cell along with the vertical air motions calculated from integration of the mass continuity equation downwards from 3.0 km using the triple Doppler value of \( w_a \) as a boundary condition. The vertical winds are apparently downwards at about -3.0 m/s in the convergence region. This compares with the upward vertical velocities that are evident in the same region of Figure 4.D.7 at 3 km altitude. Unfortunately, vertical winds calculated by integrating mass continuity upwards from the ground indicate that vertical motion is close to 0 m/s, Figure 4.D.9. While this discrepancy is disappointing, it is not surprising due to the large errors expected due to the location of the cell relative to the triple Doppler triangle (see Figure 2.B.5).

A vertical slice through the long axis of the convergence and downdraft regions, Figure 4.D.10, illustrates that the vertical velocities calculated by both methods are in qualitative agreement. This figure shows a vertical slice extending from the ground to 4 km. Upwardly integrated vertical motions show that the downdraft, with peak downward velocities of -3.2 m/s, originates at about 2 km. The downwardly integrated field shows a downdraft, originating at about 3 km, with a peak strength of about -6.0 m/s.

In Figure 4.D.8, air can be seen to be entering the convergence zone from the north. A north-south cross-section through the convergence zone, Figure 4.D.11, shows that air enters the reflectivity edge of the cloud from the north, then travels about 1 km before entering the convergence zone, which is almost completely confined within the 45 dBZ contour (as confirmed by tracing air parcel streamlines). This pattern of air entering the cell at an altitude of about 2 km and then travelling a long distance horizontally before entering the high reflectivity region and then descending is reminiscent of the flow pattern of cell F presented in Chapter 3.

The unsmoothed reflectivity field from MIT has been shown in order to illustrate that the actual reflectivity edge is indeed separated from the convergence zone. (Contoured values are always slightly smoothed to prevent a jumble of contour lines from being drawn.) Air that enters the convergence zone travels both upwards into the updraft at the south of the cell and downwards into the downdraft. Air is also entering the updraft from the south (left). While the data have been filtered near the cloud edge, as discussed above, there is
no hint of any significant convergence in the low reflectivity regions where the air has, presumably, just recently entered the cloud. Thus, quick saturation of dry entrained air, like that inferred from the convergence pattern exhibited by cell R, is not indicated.

It is parcels in the downdraft at this grid time, and 2012 Z, that impinge on the surface mesonet stations. Parcels high in the downdraft at later times, when the downdraft has deepened, are not measured at any functioning surface mesonet stations.

Similar slices through the downdraft region will be shown for the 2020 Z and 2024 Z grids. The data will show that despite inaccurate calculations, the major features of the downdraft field are persistent and that there is a trend towards a more shallow downdraft at 2024 Z.

3. 2020 Z

Four minutes later, at 2020 Z, the downdraft still appears to have origins at about 2 km above the surface. The convergence zone has shifted about 1 km to the south so that it is significantly further from the cloud edge and the entraining air. Figure 4.D.12 shows a horizontal slice at the 2 km level over the same storm-relative region as that shown for 2016 Z. Wind vectors illustrate strong flow, mostly from the northeast, into the high reflectivity region and the convergence zone of the cell. The convergence region is largely contained in a region very near the 50 dBZ contour near (60,90). Vertical wind fields, from different methods of calculation, also shown in this figure, show substantial disagreements. However, they are in qualitative agreement in the region of interest, the downdraft. The upwardly integrated vertical winds show a broad region of downward flow, peaking at about -5 m/s in several locations. Upward flow is indicated to the southwest of the convergence zone. The directly calculated then downwardly integrated vertical winds indicate that there is a much more local downdraft located in and to the west of the convergence zone. In this region, and in the updraft region to the southwest of the convergence, there is fairly close agreement with the upwardly integrated fields. The peak downward velocities in these fields are about -7 m/s.

Vertical slices through the downdraft region illustrate this consistency, but also indicate the shortcomings in the fields that are a result of inaccuracies introduced primarily by the poor location of the storm relative to the triple Doppler triangle and large inter-tilt spacing. Figure 4.D.13 is a vertical slice through the long axis of the downdraft. It shows the
convergence zone extending from 1 km to 4 km in the high reflectivity region. This slice extends east-west, so the engulfment of air from the north cannot be seen. The two vertical motion fields, while possibly appearing quite different, actually display consistency in the central regions of the cell. The updraft at the left of the plot is prominent in both fields although the upwardly integrated peak value grows with height as integration errors accumulate. At 2.6 km, at the switching level between the directly calculated and vertically integrated fields, the updraft velocities are similar: \( w_{\text{up}} = 6.5 \, \text{m/s}, \, w_{\text{dir/down}} = 5 \, \text{m/s}. \) The upwardly calculated downdraft also blows up near the top of the plot, which is a level of almost zero vertical motion according to the direct, triple Doppler, calculation. The results are similar over a large portion of the downdraft volume, however. At the southern extreme, the downdraft reaches to about 1 km in both fields and has a peak strength of either -2 m/s or -3 m/s. Near \( x = 64 \) (which will be the location of Figure 4.D.14, a north-south cross-section) the top of the downdraft is at about 2.4 km and its peak strength is \( w_{\text{up}} = -4 \, \text{m/s}, \, w_{\text{dir/down}} = -6 \, \text{m/s} \) at about 1.8 km. Both fields are unreliable at the east (right) edge of the plot due to smoothing with data that has been taken or integrated across filtered regions. The rough indication is that there is a downdraft extending from the ground to about 3 km with a likely peak strength of about -3 m/s to -6 m/s. Reference to Figure 4.C.7 will confirm that there is some small surface divergence occurring in this region at this time.

A vertical slice perpendicular to Figure 4.D.13, at \( x = 64 \), is shown in Figure 4.D.14. In this cut, the movement of air from the north (right) into the highest reflectivity region of the cloud can be seen. As the wind vectors illustrate (and traced air parcels confirm), air parcels cross the 15 dBZ contour and then travel about 2 km, which is about 300 seconds at the observed cross-cell velocity, before descending into the core of the downdraft. The wind vectors have been calculated using the directly calculated winds above 2.6 km and downwardly integrated winds below that level. The upwardly calculated fields indicate a more broadly based downdraft with indications that the air descends slowly before entering the strongest regions of the downdraft and accelerating. A comparison of the two fields, in the same manner as that applied to those of Figure 4.D.13, would show that the agreement in the regions of interest and mutual reliability is reasonable. While not prominent in this particular slice, air is entering the cell from the south (it can be seen in the horizontal cross-section, Figure 4.D.12.). This air must also travel for several minutes after crossing the 15 dBZ contour before it enters the high reflectivity and convergence region.
4. 2024 Z

At about 2024 Z, the surface outflow reaches peak strength as shown in Figures 4.C.7 and 4.C.8. The cell as a whole is well into its collapsing phase as shown in Figure 4.C.3. Reflectivities of 50 dBZ now reach to only about 2.4 km. The storm has moved westward and is now straddling the triple Doppler triangle (see Figure 4.C.9). Therefore, slightly more accurate directly calculated vertical velocities are expected. By this time the downdraft is more clearly defined as having roots near the 2 km level in a region of convergence coincident with the highest reflectivity. The peak negative velocities are about -4 m/s, though the two methods of vertical velocity calculation give somewhat different results.

A horizontal slice through the 2 km level, Figure 4.D.15, very near the top of the downdraft, illustrates the now familiar pattern of convergence in the high reflectivity region. The peak reflectivities at this altitude are now lower, 48 dBZ. There is clear evidence of an intrusion of low reflectivity into the central portions of the cell from the northeastern side (see the 40 dBZ contour). The vertical velocities at this level are small in the downdraft region, near (62,88), except for an unreliable value of -5 m/s in the directly calculated field. (It is in a region of very high three-dimensional divergence and is created from rather distant post-filtered filled data.) While there are many differences in the results of the two calculation methods, there is still qualitative agreement that is more apparent, and will be discussed, with reference to the vertical slices shown below.

An east-west slice through the downdraft, Figure 4.D.16, shows that convergence is occurring from 1 km to 3 km altitude through much of the downdraft region although there is a region to the west (left) where convergence is confined to between 800 m and 1600 m. While this shallow convergence layer is directly above the peak surface divergence visible in Figure 4.C.7, the wind vectors and calculated streamlines indicate that air parcels from this divergence region actually originate much further to the east (right). Both upwardly and downwardly integrated vertical wind fields exhibit a downdraft whose upper boundary is at about 2 km, in the convergence zone. The peak downward velocity of -6.5 m/s in the downwardly calculated field is suspect for the reasons mentioned in the previous paragraph. The residual vertical velocity at the ground is significantly negative (lowest level visible in plot is 200 m). This indicates that the downward integration is overestimating the strength of the downdraft by about 2-3 m/s. With this correction, the values are reasonably consistent with the upwardly integrated result. As discussed before,
the results of the two methods of calculation diverge at levels above 2-3 km as integration errors increasingly contaminate the upwardly integrated result. The cross-cell flow can be seen in the north-south cross-section of Figure 4.D.17. This slice is through x=58 and, as can be seen by examining Figure 4.D.15, it is through the center of the convergence zone, not the shallow western portion of the downdraft visible near (52,5) in Figure 4.D.16. In this slice, air can be seen entering the reflectivity edge of the cell from the north (right) and south (left) and then travelling into the highest reflectivity region before experiencing convergence and either descent or ascent. Air that experiences an environment exhibiting radar reflectivities of about 43 dBZ or more descends, while air entering just above this level (in lower reflectivity) ascends into the updraft.

5. 2028 Z

By 2028 Z the storm has collapsed. There is a large surface outflow occurring, both in the front and rear divergence regions, as seen in Figure 4.C.7. In the eastern portion of the downdraft, direct calculations of strength are impossible due to the shallow nature of the cell. Only one plot is shown at this final time. Figure 4.D.18 is an east-west cross-section through the outflow region. The collapse of the reflectivity core is obvious; the 40 dBZ contour reaches only as high as 2.0 km in Figure 4.C.3. The mid-level convergence region has dissipated and all that is left is a diffuse region of small (< .005 /s) convergence. While the uppermost regions of the eastern portion of the downdraft are not resolved, the upwardly integrated vertical velocity field suggests that it may extend to about 2 km. The western portion of the downdraft extends to about 3 km in both the upwardly and downwardly integrated vertical velocity fields.

In the upper reaches of the western edge of the downdraft (towards the right edge of the plotted area), the diffuse convergence does not ever completely compensate for the surface divergence and there is residual downward motion in the upwardly integrated vertical velocity field. It is not surprising to observe this imbalance during the dying stages of a downdraft, given the radar scanning strategy that has been employed. Figure 4.D.19 shows the relative times at which the individual radars scan sections of the vertical slice in Figure 4.D.18. The plotted times are relative to the nominal time used for grid shifting, 2030 Z. The region near 3 km is scanned with tilts taken 2.0, 0.8, and 1.7 minutes after the ground level tilts by the MIT, FL2 and UND radars respectively. During this time, it is likely that the diffuse convergence, noted earlier at 3 km, weakened significantly.
E. Calculations of Air Parcel Trajectories

In order to measure accelerations experienced by air parcels and to compare these accelerations to those expected based on the air parcel environment, streamlines and short trajectories were calculated.

While it is possible to calculate trajectories, and one will be shown below, the period during which air parcels tended to experience acceleration is usually shorter than the grid update time of four minutes. Since it is possible to create only short trajectories, the entire history of air parcels could not be deduced and four minute long streamline sections usually sufficed to diagnose accelerations that were usually of short duration. A typical trajectory showing a parcel entering the high reflectivity region, accelerating downwards, and then impacting the ground, is shown below in Table 4.1.

Several trajectory calculations have been made in order to trace air parcel paths in this region. Trajectories are calculated by piecing together streamlines from grids at adjacent times. This allowed the tracing of paths that are longer than the four minute inter-grid time interval. There are many problems however. Many trajectory calculations failed when the observed conditions at the end of two co-located streamline endpoints are significantly different. For example, a streamline calculated using the 2016 Z grid data ends at 2020 Z at a certain grid location that has properties of $u, v, w_a$, and $Z$. In an error free analyses, the values of these quantities in the 2020 Z grid will be identical at that point. Usually they differed significantly, primarily because of storm evolution. The speed of storm evolution is documented in the reflectivity history in figure 4.C.3. In this figure, changes of 5 dBZ can be seen at given altitudes in the four minute grid update time. While small field discrepancies can be tolerated, it makes no sense to continue a streamline if the vertical velocity changes from, say, -5 m/s to +5 m/s from one grid time to the next.
Table 4.1 Trajectory in Orlando downdraft. Various properties of an air parcel entering the downdraft are listed. Time in seconds, \((x,y)\) grid location in radar grid coordinates \((\text{one gridpoint} = 200 \text{ m})\) \((\text{Advected grid location is different at 2016 Z and 2020 Z by nine grid units following discussion in text of Section c. The endpoints of the 2016 Z and 2020 Z sections match.})\), \(z\) in km, \((u,v,wa)\) velocity, reflectivity from the MIT radar, dBZ, precipitation water content, M from Geotis (1971) Z-M relation, and change of vertical velocity during listed timestep, \(dw_a\).

The first air parcel trajectory segment begins just above \((75,95)\) in Figure 4.D.7. It follows the plotted vectors from the edge of the reflectivity into the convergence region to the southwest. It then accelerates downward in the downdraft that is visible in Figure 4.D.8. The continuation of the trajectory at 2020 Z can be followed in Figure 4.D.12. The latter section begins at \((64,88.5)\) 400 m above the 2 km altitude shown in the figure. This location is within the indicated downdraft and at the edge of the convergence region. The parcel proceeds down and to the west through this convergence and then, by inference from Figure 4.D.13, to the ground.

<table>
<thead>
<tr>
<th>Time (sec)</th>
<th>x (km)</th>
<th>y (km)</th>
<th>z (m/s)</th>
<th>u (m/s)</th>
<th>v (m/s)</th>
<th>wa (m/s)</th>
<th>dBZ</th>
<th>M (g/m(^3))</th>
<th>dw_a (m/s)</th>
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</thead>
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<td></td>
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<td></td>
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<td>.6</td>
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</table>
The trajectory shown above is comprised of two streamline segments, the first from 2016 Z data and the second from 2020 Z data. The break is indicated with the time discontinuity in the left columns. As can be seen, the parcel enters the high reflectivity region of the cell, the ambient reflectivity increasing from 41 dBZ to about 47 dBZ. During this time it accelerates downwards starting with \( w_a = 0 \) and reaching a peak of \( w_a = -6 \text{ m/s} \). This value is approximate and it also is discontinuous by about 1.5 m/s across the grid change. The air parcel experiences average accelerations of about \(-0.18 \text{ m/s}^2\). Most of the acceleration occurs after the parcel crosses into the 45 dBZ region. This value will be compared with the values predicted by simple forcing calculations below.

In order to best illustrate the fate of parcels just entering the mid and high reflectivity regions, and to examine the forcing mechanisms, a streamline, calculated using data from only the 2016 Z grid, is presented below. Its location has been chosen so that the maximum acceleration period is centered on the grid time.

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<th>x grid</th>
<th>y grid</th>
<th>z km</th>
<th>u (m/s)</th>
<th>v (m/s)</th>
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<th>( dw_a ) (m/s)</th>
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<td>47.8</td>
<td>2.10</td>
<td>0.8</td>
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</table>

Table 4.2 Streamline in Orlando downdraft. Various properties of an air parcel entering the downdraft are listed. Time in seconds, \((x,y)\) grid location in radar grid coordinates, \(z\) in km, \((u,v,w_a)\) velocity, reflectivity from the MIT radar, dBZ, precipitation water content, M, and changes in vertical velocity during listed timestep, \(dw_a\).

Figure 4.D.8 shows the streamline starting location \((70,95)\). The wind vectors indicate that air is entering the cell from the north, passing the 40 dBZ contour, and then experiencing strong horizontal convergence and descending into the downdraft near \((70,90)\). The air
parcels following this streamline originate outside the 20 dBZ region of the cell. As the parcel crosses the 41 dBZ contour it begins to accelerate downwards. Vertical accelerations experienced by parcels along this streamline are approximately .025 m/s².

While complete trajectories extending from outside the cell, into the downdraft, and down into the outflow could not be constructed, piecewise, somewhat discontinuous trajectories that were examined subjectively indicate that air enters the convergence region and descends in the downdraft to the outflow regions observed at the ground (observed by radar, that is, since no functioning mesonet stations experienced the central portions of the outflow).

F. Forcing Mechanisms of the Orlando Downdraft

Simple calculations of the magnitude of various forcing mechanisms are shown in Section 1. Discussion of one-dimensional model simulations are presented in Section 2.

1. Calculations of Instantaneous Forcing of the Orlando Downdraft

The convergence and acceleration regions in this downdraft are well within the region of detectable reflectivity and hence probably (see Chapter 1, Section G) well within the cloud boundary. The reflectivity edge is at least 1 km and usually about 2 km from the convergence region, as seen in Figure 4.D.8, for example. As discussed before with respect to cell F, this implies that instantaneous cooling by the entrainment of subsaturated environmental air is not a likely acceleration mechanism here.

The top of the downdraft has been shown to be between 2 km and 3 km altitude, see Figures 4.D.7, 4.D.8, and 4.D.10, which is well below the environmental melting level of 4.5 km. Therefore cooling due to melting ice is not a likely acceleration mechanism in this case.

The precipitation water contents that are inferred to be present in the downdraft region, about 2 g/m³ (using the Geotis (1971) Z-M relation), are consistent with mass loading accelerations of about .02 m²/s², following Table R.3 in Chapter 3.
2. Model Simulation of the Orlando Downdraft

The one-dimensional model was run to simulate the downdraft. The model was initialized using saturated, in cloud air at the top of the downdraft. This was consistent with the observations discussed above and shown in Figures 4.D.7, 4.D.8 and 4.D.4b, among others. Specifically, since the model is initialized at the top of the downdraft, the initial air parcels have a long residence time in the moderate reflectivity regions of the cell. Thus, they will most likely have been saturated due to rain drop and cloud droplet evaporation as they mixed with cloudy air before entering the convergence and acceleration region. The model was initialized with the following values of variables:

<table>
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<th>Variable</th>
<th>Value</th>
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<tbody>
<tr>
<td>Radar reflectivity dBZ</td>
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</tr>
<tr>
<td>Cloud droplet concentration in initial parcel</td>
<td>0 g/m³</td>
</tr>
<tr>
<td>Ice particle concentration in cloud</td>
<td>0 g/m³</td>
</tr>
<tr>
<td>Entrainment parameter</td>
<td>α = .25</td>
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<tr>
<td>Initial radius r</td>
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<tr>
<td>Pressure level of simulated boundary level PBL</td>
<td>960 mb</td>
</tr>
<tr>
<td>Pressure level of ground PGR</td>
<td>1020 mb</td>
</tr>
<tr>
<td>Cloud droplet concentration in cloudy air ρyc</td>
<td>1.0 g/m³</td>
</tr>
</tbody>
</table>

The model soundings, were in (pressure(mb), temperature(K), relative humidity) format:

- **outside environment**
  - (700,280,.95)
  - (800,288,.75)
  - (1020,303,.62)
- **cloudy entrainment environment**
  - (700,281,1.)
  - (910,293,1.)
  - (1000,303,.62)

These terms and the one-dimensional model are described in Appendix 2.

The model sounding that is presented was derived subjectively from the soundings presented in Section B. It was not modified after its initial determination in so as to avoid the model tuning morass. The presented model results use initial and environment variables that were determined prior to the first simulation to preserve objectivity.
The model downdraft reaches a peak strength of about -7 m/s. This is slightly higher than most radar calculations of the downdraft velocity, the \(-6 \text{ m/s}\) in Figure 4.D.10 for example, but some calculations did result in \(-7 \text{ m/s}\) so this is at least a consistent result. The thermodynamic qualities of the model downdraft compare reasonably well with the observations taken at the mesonet stations and with the Orlando airport observations. Unfortunately, as can be seen in Figure 4.C.7, station 23 is affected very early in the outflow history, when the downdraft is shallow and weak, and station 22 is only glancingly hit. Therefore the mesonet observations do not place as firm a constraint on the model results as did those in Huntsville. In addition, the lack of a balloon sounding that is proximate in time or space to the event, raises doubts about the representativeness of the sounding used to produce the model outflow. The model downdraft is quite sensitive to perturbations in the environmental sounding, consistent with the results of Proctor (1989).

The simulations produce surface θ_e values of just over 348 K. This is 4 K less than the mesonet observations and 2 K less than at Orlando airport. However, this is not inconsistent with the surface mesonet observations in which θ_e did not change significantly during the event. First, a four degree θ_e drop is not very large and second, maybe it did occur at the surface directly under the downdraft and was not observed. Parcels reaching the mesonet stations had time to mix with surface air before being measured.

The model downdraft reaches the ground too warm (model T = 301 K, observations T = 298 K) and dry (model RH = .71, observations RH = .95) for reasons that have been discussed in detail in Chapter 3.

The dominant forcing mechanism appears to be precipitation loading. The precise magnitude of the forcing is strongly dependent on the precipitation water content, which, in

<table>
<thead>
<tr>
<th>p(mb)</th>
<th>z(km) approx</th>
<th>θ_e (K)</th>
<th>T(K)</th>
<th>RH</th>
<th>W (m/s)</th>
<th>B_T (m/s^2)</th>
<th>B_L (m/s^2)</th>
<th>ρ_v (g/m^3)</th>
</tr>
</thead>
<tbody>
<tr>
<td>800</td>
<td>2.2</td>
<td>347</td>
<td>288</td>
<td>1.00</td>
<td>0</td>
<td>.02</td>
<td>-.031</td>
<td>12.7</td>
</tr>
<tr>
<td>900</td>
<td>1.2</td>
<td>346</td>
<td>294</td>
<td>.84</td>
<td>6</td>
<td>-.016</td>
<td>-.027</td>
<td>15.1</td>
</tr>
<tr>
<td>950</td>
<td>0.7</td>
<td>347</td>
<td>297</td>
<td>.74</td>
<td>7</td>
<td>-.017</td>
<td>-.025</td>
<td>16.6</td>
</tr>
<tr>
<td>1000</td>
<td>0.2</td>
<td>348</td>
<td>301</td>
<td>.71</td>
<td>0</td>
<td>0</td>
<td>0</td>
<td>19.2</td>
</tr>
</tbody>
</table>

Table 4.3 Model simulation of Orlando downdraft. \(z\) is altitude, \(p\) is pressure, \(T\) is temperature, \(RH\) is relative humidity, \(W\) is the downdraft vertical velocity, \(B_T\) is the thermal buoyancy acceleration, \(B_L\) is the precipitation loading acceleration, \(ρ_v\) is the vapor density.
turn is very sensitive to the observed reflectivity. The values used in the model simulation were typical of the observations in the downdraft region. As can be seen in Table 4.3, the mass loading, $B_L = -0.031 \text{ m/s}^2$ is only marginally overcoming the positive buoyancy of the air, $B_T = 0.020 \text{ m/s}^2$ at the top of the downdraft. This explains why the convergence and acceleration zone is confined to the highest reflectivity regions of the cell within the 45-50 dBZ contours. This behavior is similar to that noted for cell F in Huntsville as described in Chapter 3.

As is the case with the other downdrafts, evaporation of precipitation is crucial to the sustenance of the downdraft. (There is probably not much cloud water content in the lower portions of the cell and it is quickly depleted in descending and warming parcels in the downdraft.) The evaporation into the descending parcels cools the parcels relative to dry adiabatic values. Eventually, below 890 mb, just above cloud base, the downdraft becomes slightly downwardly buoyant. This thermal buoyancy is never larger than the mass loading acceleration, however.

Since the observed downdraft is well within the cell and is observed to be entraining air that has long in-cloud residence times, the model downdraft entrains warm cloudy air during descent. The evaporation of the incorporated cloud droplets produces 2.1 degrees of cooling to the simulated downdraft, compared with 3.7 degrees from the evaporation of raindrops. This is a significant fraction but is not absolutely necessary for the maintenance of the downdraft. Model simulations with no cloud droplets produce a very similar downdraft.

A model simulation was conducted using initially subsaturated air at 3 km in order to simulate the possible effects of the nearby southern edge of the cloud. The results were similar except that the surface $\theta_e$ is significantly lower. It is tempting to exclude this as inconsistent with the mesonet observations of constant $\theta_e$, but, as noted earlier, the observations are not as constraining in this case as would be desirable.

G. Summary

Multiple Doppler analyses have shown that the downdraft in the Orlando cell has its origins at low- to mid-levels, 2 km - 3 km, in the small storm. Trajectories and streamlines indicate that air parcels travelled for some time inside the cell before being drawn.
downwards by the weight of precipitation in the high reflectivity regions of the storm. Maximum observed downward velocities are about - 6 m/s.

The peak surface outflow is well observed by the radars but is only poorly sampled by the mesonet stations. The mesonet observations indicate that the thermodynamic characteristics of the outflow air are similar to that of Huntsville cell F, described in Chapter 3. The outflow air exhibits high values of $\theta_e$ and $\rho_v$. $\theta_e$ is near pre-event levels and $\rho_v$ increases significantly.

Simple model simulations, initialized with crude interpolated soundings, agree reasonably well with the observed radar characteristics of the downdraft. The model produces an outflow with slightly depressed values of $\theta_e$. These values are consistent with the observed lack of low $\theta_e$ air at the mesonet stations, although this is a weak constraint due to the distance of the stations from the peak outflow.

As with the Huntsville downdrafts, the dominant forcing mechanisms are the combination of precipitation loading providing downward acceleration and evaporation of raindrops keeping the descending parcels reasonably cool. Some evaporation of cloud droplets is probably occurring but is not a necessary factor in producing the downdraft.
CHAPTER 5 SUMMARY

A. Downdraft Forcing Mechanisms

Four downdrafts have been studied with multiple Doppler radar analyses and surface mesonet observations. While the downdrafts exhibit differing origin heights and strengths, the dominant forcing mechanisms in all cases are precipitation loading and evaporation of rain. The structures of the downdrafts, the characteristics of the thermodynamic observations at the surface and the roles of various possible forcing mechanisms are summarized in Table 5.1.

<table>
<thead>
<tr>
<th>Downdraft</th>
<th>Depth km</th>
<th>Surf θg drop</th>
<th>Ice present</th>
<th>Dominant Forcing</th>
<th>Secondary Forcing</th>
<th>Wmin m/s</th>
<th>ΔV m/s</th>
<th>dBZ</th>
<th>Z40 km</th>
</tr>
</thead>
<tbody>
<tr>
<td>Cell R Huntsville</td>
<td>4-5</td>
<td>8 K small</td>
<td></td>
<td>Precip Loading In-Cell Rain Evap</td>
<td>Cloud Droplet Evaporation</td>
<td>6</td>
<td>14</td>
<td>50</td>
<td>8.8</td>
</tr>
<tr>
<td>Cell F Huntsville</td>
<td>2</td>
<td>no no</td>
<td></td>
<td>Precip Loading In-Cell Rain Evap</td>
<td>Below-Cell Rain Evaporation</td>
<td>9</td>
<td>17</td>
<td>55</td>
<td>8.3</td>
</tr>
<tr>
<td>Cell S early Huntsville</td>
<td>2</td>
<td>no no</td>
<td></td>
<td>Precip Loading In-Cell Rain Evap</td>
<td>Below-Cell Rain Evaporation</td>
<td>4</td>
<td>12</td>
<td>55</td>
<td>6-7</td>
</tr>
<tr>
<td>Cell S late Huntsville</td>
<td>5</td>
<td>8 K yes</td>
<td>yes</td>
<td>Precip Loading In-Cell Rain Evap</td>
<td>Cloud Dr Evap Melting</td>
<td>8-10</td>
<td>24</td>
<td>60</td>
<td>&gt;8</td>
</tr>
<tr>
<td>Cell S late front Huntsville</td>
<td>2</td>
<td>no no</td>
<td></td>
<td>Precip Loading In-Cell Rain Evap</td>
<td>Below-Cell Rain Evaporation</td>
<td>5</td>
<td>--</td>
<td>60</td>
<td></td>
</tr>
<tr>
<td>Cell O Orlando</td>
<td>2-3</td>
<td>no no</td>
<td></td>
<td>Precip Loading In-Cell Rain Evap</td>
<td>Below-Cell Rain Evap, Cloud Evap</td>
<td>6</td>
<td>15</td>
<td>50</td>
<td>5.8</td>
</tr>
<tr>
<td>Kingsmill Huntsville</td>
<td>4</td>
<td>--- yes</td>
<td></td>
<td>Precip Loading then Below-cell Rain Evap, and Melting</td>
<td></td>
<td>12</td>
<td></td>
<td>&gt;50</td>
<td></td>
</tr>
<tr>
<td>Mahoney #1 Denver</td>
<td>2-3</td>
<td>--- assumed</td>
<td></td>
<td>various speculated</td>
<td></td>
<td>7</td>
<td>&gt;40</td>
<td></td>
<td></td>
</tr>
<tr>
<td>Mahoney #2 Denver</td>
<td>3-5</td>
<td>6 K assumed</td>
<td></td>
<td>various speculated</td>
<td></td>
<td>10</td>
<td>&gt;40</td>
<td></td>
<td></td>
</tr>
</tbody>
</table>

Table 5.1 Summary of Downdraft Results. Downdrafts in this study labelled according to text. Kingsmill from Kingsmill and Wakimoto (1991) and Tuttle et al. (1989). Mahoney from Mahoney and Elmore (1991). These authors did not quantify forcing terms so primary vs. secondary mechanisms were difficult to determine. ΔV is max differential radial velocity across outflow from radar. dBZ is maximum storm reflectivity. Wmin is peak downward velocity in downdraft. Z40 is highest penetration of 40 dBZ contour.
While the forcing mechanisms in the four studied cases are similar, a generalization of these results to all downdrafts would be unwise. While it is shown here that a deep downdraft can form without ice, it is unclear from this limited sample just what fraction actually do. Clearly more studies of moist sub-cloud environment, low cloud base storms need to be conducted.

The environments of the studied downdrafts, containing moist and shallow sub-cloud layers, is typical of the Eastern United States. Conditions are frequently very different in the Plains and Western United States. Sub-cloud evaporation of rain, as discussed by Srivastava (1985) is probably much more significant in the dry and deep sub-cloud environments typical of those regions. The sub-cloud layers under the cells studied here are very shallow, only about 1 km in thickness. Downdrafts do not have much opportunity to accelerate in these layers, whatever the forcing potential, before entering the surface boundary level and decelerating. The deceleration layer, as diagnosed by radar observations of near-surface divergence (see Figures R.16, F.13, F.18, S.14, S.20, 4.D.11, 4.D.16, and 4.D.18), extends from the ground to between 400 m (Figure 4.D.18) and 1000 m (Figures R.16, F.18, S.20) altitude. These observations are consistent with those from a large number of outflow events in Huntsville, shown in Figure 5.1. In the figure, Foreman (pers. comm.) shows that the depth of surface outflows is typically 400 m - 1000 m. Since cloud bases were near 1000 m altitude in the downdrafts studied here, the region in which downward forcing mechanisms could act was limited to between 600 m and zero meters.

Precipitation loading, important in all of the high reflectivity storms that were studied here, may be less important in the low reflectivity-low precipitation water content events that are common in the Western United States (see Figure 1.9), though the strongest outflows in those areas frequently occur in concert with high reflectivities.

The proximity soundings in Huntsville indicate that low $\theta_e$ air is confined to mid-levels on most days during which downdrafts are observed. (The median 1000 mb - 900 mb $\Delta \theta_e$ was 5 K and the median 1000 mb - 800 mb $\Delta \theta_e$ was 13 K during the summer of 1986.) Therefore, the large outflow associated surface $\theta_e$ drops that are common in Huntsville (see Figure 1.4) imply that deep downdrafts, like those in cell S and cell R, or downdrafts containing ice, are frequent. The smaller average $\theta_e$ drops that are observed in Denver (see Figure 1.4) suggest that sub-cloud, Srivastava (1985) events may
predominate. Outflows exhibiting small $\theta_e$ drops are also fairly common in Huntsville, indicating that either Srivastava (1985) type downdrafts or somewhat deeper downdrafts, similar to cell F and cell O, also occur frequently.
APPENDIX 1  \( \theta_e \) Jumps at Mesonet Stations

Observations from several surface mesonet stations indicate that the value of \( \theta_e \) rises sharply at the onset of outflow events. Figure S.6 (station 115) shows the most extreme example of this effect that was noticed during this study. At 2034 Z, just as the relative humidity rises and the temperature drops, the value of \( \theta_e \) jumps from 354 K to 362 K.

Other examples are found in Figures S.6 (station 112 (6 K), station 114 (3 K), station 33 (1 K), station 37 (2 K)), Figure R.7 (station 104 (5 K)), Figure R.8 (station 24 (2 K)), Figure 4.C.10 (station 22 (3 K)), and Figure 4.C.11 (station 23 (3 K)).

The rises in \( \theta_e \) are coincident with increases in wind speed. One speculation is that these rises are due to the increased mixing of air from very near the ground up to thermometer level. The thermometer level in the Lincoln Laboratory mesonet stations (enumerated 100-130 in Figure R.6) was approximately 2 m (based on a photograph in Wolfson et al. (1987), see Wolfson and Iacono (1987) for a discussion and comparison of Lincoln Laboratory and PAM type stations). The level in the PAM mesonet stations (enumerated 1-41) was similar (based on a photograph in Brock et al., 1986).

Geiger (1965) presents observations from Rossi (1933) and Franssila (1936) reproduced in Figure A1.1, which illustrate the higher water vapor pressures that are typical of air near the ground. Geiger (1965) also presents data from Brocks (1948), Figure A1.1, that indicate that lapse rates of temperature approach 70 K/100 m at 1 m altitude. Tabular data from Best (1935), also in Geiger (1965), show that lapse rates can approach 77 K/100 m in the 0.3 m to 1.2 m layer and 682 K/100 m in the 0.025 m - 0.3 m layer.

The presented data, taken in England and Finland, may represent environments that are not characteristic of the Southeastern United States. At the temperatures common to Huntsville and Orlando, over 303 K during summer afternoons, the temperature lapse rates from Brocks (1948) would result in increases in surface water vapor pressure of the magnitude observed by Rossi (1933) and Franssila (1936) if the relative humidity was constant.*

Based on the lapse rate observations of Brocks (1948) and Best (1935) it is possible that the air in the lowest meter of the atmosphere was about 0.5 K warmer than that at the

* The saturation water vapor pressure, \( e_{sv} \), is a steep function of temperature at high temperatures, so that \( e_{sv}(303.2 \text{ K}) = 42.6 \text{ mb} \) and \( e_{sv}(304.2 \text{ K}) = 45.1 \text{ mb} \). Thus, an increase of one degree causes an increase in water vapor pressure (at constant relative humidity) of 6%.
observation level of 2 m. Based on the water vapor observations of Rossi (1933) it is possible that the water vapor density was as much as 5% higher at 1 m that at 2 m. While air in the lowest 0.5 m of the atmosphere was both warmer and moister, the small quantity of air in this thin layer will be diluted by any plausible process that mixes it to at least the observation height.

If the air temperature increased by 0.5 K due to increased mixing of ground level air to thermometer height, the resultant increase in \( \theta_e \) would be about 2 K. This is smaller than the observed jumps, particularly at station 115, and is based on the very generous assumption that the hot, moist near-surface air is not mixed through a layer that has an upper boundary well above the observation height. (A deeper mixing layer would cause a dilution of the effect.) This effect may explain the gradual rise and sustained increases in \( \theta_e \) observed at the mesonet stations in Orlando. These stations (see Figures 4.C.10 and 4.C.11) experienced increased winds as they were grazed by an outflow event. The \( \theta_e \) rises were small and the values did not immediately drop after the event.

In Huntsville, the \( \theta_e \) jumps are systematically larger at the Lincoln Laboratory mesonet stations than at the PAM mesonet stations. This leads to the speculation that the jumps may be due to instrumental effects. The jumps in \( \theta_e \) occur during periods when the temperature and relative humidity are changing rapidly (At station 115, the temperature drops from 303 K to 299 K within four minutes while the relative humidity increases from .61 to .92.) DiStefano (pers. comm.) claims that the Lincoln Laboratory mesonet stations contain thermometers which achieve a 65% response to changes within 30 seconds and humidity sensors that achieve a 90% response within 15 seconds. These different response times can cause jumps in observed values of \( \theta_e \). Figure A1.2 shows the calculated instrumental response to an instantaneous change in temperature and relative humidity during which \( \theta_e \) is conserved. The calculated value of \( \theta_e \) jumps by almost 9 K for a short period. This is more than the peak value of 7 K that is exhibited by station 115. Also shown are calculations of \( \theta_e \) based on simulated values of the one minute average values of temperature and relative humidity that the mesonet stations actually report. The time period over which the averages has been taken has been shifted by 0, 10, and 20 seconds relative to the time of the saturation event to illustrate the effect of an event impacting on a station during the middle of one of the one minute averaging periods.

While the magnitude of the initial \( \theta_e \) jumps is explained, the extended duration of the increased \( \theta_e \) values are not. At station 115 the values of \( \theta_e \) are increased by 4 K, 8 K, and
3 K compared to the pre-outflow levels during the first three minutes of the outflow event. The value of $\theta_e$ that would be observed during the third minute of an event is only increased by up to 1.5 K in the calculations shown in Figure A1.2. It is possible that the increased $\theta_e$ values observed during the later period are due to wetting of the humidity sensor that is discussed further in Appendix 3. It is also possible that the instrumental response times are not as fast as reported or that the mixing effect discussed above also contributes.
APPENDIX 2 Description of one-dimensional Model

A. Overview

A one-dimensional model was developed to permit the diagnoses of the relative importance of specific forcing mechanisms in the radar and mesonet observed downdrafts. The model includes detailed thermodynamic calculations that simulate the roles of raindrops, cloud droplets, wet ice particles and dry ice (water) particles. Conversions between the various types of particles and between the particles and water vapor are allowed through evaporation, sublimation, deposition, condensation, melting and freezing. Entrainment of either in-cloud or out-of-cloud air is simulated. Thermal buoyancy, water loading and entrainment drag forces act to accelerate the simulated parcels. Model parameters are highly constrained by the radar, balloon, and surface measurements in and near the actual downdraft being simulated.

As seen in Chapter 3 and Chapter 4, the model accurately simulates the observed downdrafts. This correspondence is achieved with no adjustments or model tuning and thus inspires confidence in the resultant diagnosed forcing terms.

B. Motivation

Crude estimates of the importance of various proposed mechanisms of downdraft forcing are presented in Chapter 1 and Chapter 3. Those calculations and the observations tend to indicate that, under favorable circumstances, multiple mechanisms may be acting simultaneously. The crude calculations do not attempt to account accurately for parcel movement, the time scales associated with some of the mechanisms, or the interaction of mechanisms with one another.

The model was developed with the sole intent of diagnosing forcing mechanisms. This was not an attempt to comprehensively simulate the three-dimensional and temporal evolution of the downdrafts. The intent was to duplicate the observed properties of particular observed downdrafts and diagnose non-observed quantities such as forcing mechanisms. The model could be constrained by using the very detailed observations of the downdrafts presented in Chapter 3 and Chapter 4.
C. Model Physics

The model downdraft has the following properties at any particular level:

<table>
<thead>
<tr>
<th>Property</th>
<th>Description</th>
</tr>
</thead>
<tbody>
<tr>
<td>Pressure</td>
<td>Water Content held at 273 K in association with Ice</td>
</tr>
<tr>
<td>Water Vapor Content</td>
<td>Ice Content</td>
</tr>
<tr>
<td>Cloud Droplet Content</td>
<td>Vertical Velocity</td>
</tr>
<tr>
<td>Raindrop Content</td>
<td>Horizontal Scale</td>
</tr>
</tbody>
</table>

Processes that act to modify these properties are illustrated in the schematic diagram of Figure A2.1, where various model processes have been labelled.

1. Conversion Between Raindrops and Vapor: Evaporation and Condensation

A parameterization of raindrop evaporation and condensation has been developed. Other workers, Srivastava (1985) for example, have calculated evaporation rates for different size raindrops and tested the results of simulating different raindrop size distributions. The parameterization of evaporation over the entire raindrop spectrum has been conducted by Klemp and Wilhelmson (1978), Tripoli and Cotton (1980), Kessler (1969), and Orville and Kopp (1977) among others. The parameterization that are used by these workers include empirical formulations that are difficult to root in simple physics. The different parameterizations also result in different evaporation rates when applied to identical environments, as will be shown below. A derivation of an evaporation scheme using only simple and observed properties of raindrops is presented here.

From Pruppacher and Klett (1980) the rate of change of mass of a single raindrop due to evaporation is:

\[
\frac{dm}{dt} = \frac{d^4\pi a^3 \rho_w}{dt} = 4\pi a D_v \Delta \rho_v
\]

(1)

where \( m \) is the mass of the drop, \( t \) is time, \( a \) is the radius of the drop, \( \rho_w \) is the density of liquid water, \( D_v \) is the diffusion coefficient for water vapor in air* and \( \Delta \rho_v \) is the difference

* Following Hall and Pruppacher (1976), \( D_v = 2.11 \times 10^{-5} \ (T/273.15)^{1.94} (1013.25/P) \ {m}^2/{s} \), \( T \) in K, \( P \) in mb
in vapor density between the air at the drop surface and the air at a great distance from the drop.

Raindrops are not stationary with respect to the air in which they reside, of course. The motion of raindrops through the air enhances the purely diffusional evaporational process described in (1). An empirical ventilation coefficient, \( F \), is often defined to account for this effect:

\[
\frac{dm}{dt} \text{(ventilated)} = F \frac{dm}{dt} \text{(unventilated)}
\]  

(2)

Pruppacher and Klett (1980) present a formulation for \( F \) that depends on fractional powers of the Schmidt and Reynolds numbers* and compares favorably with actual data from Kinzer and Gunn (1951). In order to find an analytic solution for the evaporation rate of a spectrum of ventilated raindrops, it is desirable to find a simpler relation for \( F \).

Figure A2.2 shows the ventilation factor \( F \) as a function of raindrop radius according to formulations presented in Beard and Pruppacher (1971) and Kinzer and Gunn (1951). Two linear relations, \( F = 10a \) and \( F = 7.5a \) are proposed here and plotted. The linear relations closely approximate the more complex formulations. The \( F=10a \) line produces values of \( F \) that are within 10% of the Beard and Pruppacher (1971) results between 0.3 mm and 1.2 mm. Above 1.2 mm, it overestimates the ventilation factor and below 0.3 mm it produces values that are too low. The linear relations predict \( f(a=0) = 0 \) where, obviously, \( f(a=0) \) should be equal to one.

To evaluate the impact of using the linear drop ventilation formulae on evaporation rates in air parcels, a raindrop size distribution must be assumed. By combining the Z-M and Z-R relations presented in Geotis (1971) and discussed in Chapter 2,

\[
Z = 21000 M^{1.43} \quad \text{where } Z \text{ is in } \text{mm}^6/\text{m}^3 \text{ and } M \text{ is in } \text{g/m}^3
\]  

(3)

\[
Z = 400 R^{1.3} \quad \text{where } Z \text{ is in } \text{mm}^6/\text{m}^3 \text{ and } R \text{ is in } \text{mm/hr},
\]  

(4)

* The Schmidt number is the ratio of the kinematic viscosity to the diffusion coefficient of a particular particle so that \( N_{sc} = \nu/D \), where \( \nu \) is the kinematic viscosity and \( D \) is the diffusion coefficient. The Reynolds number is the ratio of inertial to viscous forces \( N_{re} = UL/\nu \), where \( U \) and \( L \) are characteristic velocity and distance scales. Pruppacher and Klett (1980) present a formula for raindrop ventilation that is linearly dependent on \( N_{sc}^{1/3} N_{re}^{1/2} \) .
a distribution similar to the Marshall and Palmer (1948) form can be calculated:

\[ N(a) = N_0 e^{-2\lambda a} \]  
\[ \lambda = 4.1 R^{-0.21} = 2.163 M^{-0.231}. \]  

The evaporation rate per unit volume of air containing a distribution of raindrops consistent with (5) and (6) can be calculated from (2) using the Beard and Pruppacher (1971) and the linear formulations for F. The results for low, medium and high liquid water contents as a function of raindrop size are shown in Figure A2.3. The values of raindrop water content, 0.5 g/m³, 3 g/m³, and 6 g/m³, correspond to 39 dBZ, 50 dBZ, and 56 dBZ radar reflectivities according to Geotis (1971).

The calculations show that the linear F=10a relation is an excellent approximation to the Beard and Pruppacher (1971) results over a wide range of M. At the lowest raindrop water contents, a relation of F=11a more closely approximates the data, but the F=10a formulation is still a good estimate. It is interesting to note that most evaporation occurs from drops with radii less than 1 mm and that the peak contribution to total evaporation comes from drops with radii near 0.4 mm (raindrop concentration 0.5 g/m³) to 0.6-0.7 mm (raindrop concentration 6 g/m³). The difference between evaporation rates integrated over all drop sizes as predicted by Beard and Pruppacher (1971) and the F=10a formulation are shown below (along with the F=11a formulation for M=0.5 g/m³).

<table>
<thead>
<tr>
<th>M g/m³</th>
<th>Δ evaporation rate F=10a</th>
<th>Δ evaporation rate F=11a</th>
</tr>
</thead>
<tbody>
<tr>
<td>0.5</td>
<td>-6.4 %</td>
<td>+3.0%</td>
</tr>
<tr>
<td>3</td>
<td>+1.8%</td>
<td></td>
</tr>
<tr>
<td>6</td>
<td>+5.2%</td>
<td></td>
</tr>
</tbody>
</table>

Table A2.1 Evaporation rate differences that arise between linear ventilation formulations and the formulation of Beard and Pruppacher (1971) which closely approximates measured values from Kinzer and Gunn (1951). M is raindrop water content.
In the model formulation, (1) and (2) are combined using $F = f a$, where $f$ is a constant, to produce:

$$\frac{da}{dt} = \frac{D_v \Delta \rho_v f}{\rho_w}$$

(7)

The raindrop water content in a given volume of air before any evaporation has taken place can be written as:

$$M(t=0) = \int_{a_{\min}}^{a_{\max}} \frac{4}{3} \pi a^3 \rho_w N(a) \, da = \int_{a_{\min}}^{a_{\max}} \frac{4}{3} \pi a^3 \rho_w N_0 e^{-2\lambda a} \, da$$

(8)

By substituting (7) into (8), the raindrop water content as a function of time can be determined.

$$M(t) = \int \frac{4}{3} \pi a^3 N_0 e^{-2\lambda a} e^{-2\lambda} \frac{D_v \Delta \rho_v f}{\rho_w} t \, da$$

(9)

Thus, the raindrop water content decreases exponentially with time as long as $\Delta \rho$ is constant:

$$M(t) = \exp\left(-2\lambda \frac{D_v \Delta \rho_v f}{\rho_w} t\right) M(t=0).$$

This condition is not normally realized, of course, and the model calculates $\frac{dM}{dt}$ during each timestep using the appropriate value of $\Delta \rho$ and $M$.

$$\frac{dM}{dt} = -2\lambda \frac{D_v \Delta \rho_v f}{\rho_w} M(t)$$

(10)

The determination of $\Delta \rho_v$ is not straightforward. This term is dependent on the value of the water vapor density far from the drop, which is easily determined from the temperature and relative humidity, and the saturation water vapor density at the surface of the drop, the determination of which is more problematical since it depends on raindrop temperature.
There are different methods of determining the raindrop temperature and since the Clausius-Clapeyron relation can be a steep function of temperature, these result in moderately different values of the saturation vapor density and $\Delta \rho_v$.

Johnson (1950) presents laboratory data showing that the thermal balance condition

$$T_{\text{air}} - T_{\text{drop}} = 1.15 \left( \frac{D_v L}{\kappa} \right) \Delta \rho_v,$$

where $\kappa$ is the thermal diffusivity of air and $L$ is the latent heat of vaporization, is obeyed. Beard and Pruppacher (1971) present essentially the same relation, based on wind tunnel observations, except that the multiplicative factor on the right hand side of the equation is the ratio of the Sherwood and Nusselt numbers which is typically about $1.04$. Ranz and Marshall (1952a,b) present measurements of drop temperatures for drops evaporating in what is described as "dry" air. Watt (1971) presents the relation

$$T_{\text{drop}} = T_{\text{air}} \left[ 1 - \frac{\beta(1-\varphi)}{1+\alpha} \right] \text{ where } \alpha = \frac{L}{R_v T_{\text{air}}}, \beta = \frac{D_v L \rho_{sv}(T_{\text{air}})}{\kappa T_{\text{air}}}, \varphi = \frac{\rho_v(T_{\text{air}})}{\rho_{sv}(T_{\text{air}})}$$

where $R_v$ is the gas constant for water vapor and $\rho_{sv}$ is the saturation vapor density of water in air.

Unfortunately, these formulae and data can differ significantly as shown in Figure A2.4.* The figure shows the predicted drop temperature according to the Beard and Pruppacher (1971) and the Watt formulae and the graphical formula actually presented in the Watt (1971) paper, all calculated assuming $T_{\text{air}} = 293.15$ K and pressure $= 1000$ mb. The Watt (1971) graphical data agree fairly well, within $0.6$ K, with the Beard and Pruppacher (1971) formula at relative humidities above 0.5. The Watt (1971) formula,

---

* In Figure A2.4 and in the model calculations,

$$\kappa = 0.024 + 0.0008(T-273.15) \text{ m s K}^{-1}$$

$T_{\text{wet bulb}}$ = the temperature at which saturated air would have the same $\theta_e$ as the actual temperature so that $\theta_e(T, \text{Relative humidity}) = \theta_e(T_{\text{wet bulb}}, \text{Relative humidity}=1)$

$T_{\text{sat}} = 2501000 - 2370(T-273.15)$ J/kg with $T$ in K

$\rho_{sv}$ is calculated from saturation vapor pressure $e_{sv} = 6.112 \exp \left( \frac{17.67(T-273.15)}{T-273.15+243.5} \right)$ with $T$ in K and $e_{sv}$ in mb
however, predicts a drop temperature that is lower than the Beard and Pruppacher (1971) formula by over 1 K at relative humidities below 0.65. The drop temperature is more than 2.5 K lower at relative humidities below 0.5. At lower relative humidities (not plotted) the wet bulb and Beard and Pruppacher (1971) formulae differ by about 0.5 K while the Watt (1971) formula predicts drop temperatures that are 26.5 K lower than the wet bulb temperature at zero relative humidity.

The result of these differences is that evaporation from drops that are assumed to be at the wet bulb temperature is faster than from drops at temperatures calculated using the other methods. The magnitude of these differences is illustrated below in Table A2.2. As can be seen, the evaporation rates are in qualitative agreement (within 10%) with the exception of the Watt (1971) formula, which predicts much lower values.

<table>
<thead>
<tr>
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</thead>
<tbody>
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<td>1.2</td>
<td>1.1</td>
<td>0.8</td>
<td>1.1</td>
</tr>
</tbody>
</table>

Table A2.2 Evaporation rates under different assumed raindrop temperatures. Rates are in arbitrary units.

Ranz and Marshall (1952a,b) presented results from several measurements of water drop temperature where the air temperature was approximately 297 K - 298 K and the pressure was 990 mb. The drops were exposed to "dry" air which had a relative humidity of, presumably, nearly zero. The drop temperatures were in the range of 280.5 K to 281.0 K which was about 0.5 K warmer than the wet bulb temperature and 1 K warmer than the prediction of the Beard and Pruppacher (1971) formula. This result implies that the Watt (1971) formula is not to be trusted, particularly at low relative humidity. Since the experimental results are close to the wet bulb temperature and to the Beard and Pruppacher (1971) formula the use of either formula should produce reasonable results.

Comparisons were made between the evaporation rates predicted here and the rates predicted by Klemp and Wilhelmson (1978), Ogura and Takahashi (1971), Orville and Kopp (1977), and Tripoli and Cotton (1980). Some of the equations that are presented in the literature are incorrect as printed or do not reproduce the presented numerical results.
(Kessler (1969), Ogura and Takahashi (1971)). The results of the comparisons are shown below in Table A2.3.

<table>
<thead>
<tr>
<th>$T_{\text{air}}$ K</th>
<th>277.2</th>
<th>277.2</th>
<th>277.2</th>
<th>277.2</th>
<th>277.2</th>
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</thead>
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<td>1000</td>
<td>1000</td>
<td>1000</td>
<td>1000</td>
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<td>Relative Humidity</td>
<td>.8</td>
<td>.95</td>
<td>.5</td>
<td>.8</td>
<td>.8</td>
<td>.8</td>
<td>.8</td>
</tr>
<tr>
<td>Raindrop water content g/m³</td>
<td>3</td>
<td>3</td>
<td>3</td>
<td>1</td>
<td>6</td>
<td>3</td>
<td>3</td>
</tr>
</tbody>
</table>

| Equat (10) $T_{\text{drop}}=T_{\text{wet bulb}}$ | 1.56 | .39 | 3.99 | .67 | 2.68 | 3.70 | 1.90 |
| Equat (10) $T_{\text{drop}}=T_{\text{Beard}+\text{Pruppacher}}$ (1971) | 1.45 | .37 | 3.74 | .63 | 2.49 | 3.01 | 1.73 |
| Equat (10) $T_{\text{drop}}=T_{\text{Wat}}$ (1971) | 1.37 | .36 | 3.09 | .59 | 2.34 | 2.04 | 1.56 |
| Tripoli and Cotton (1980) | 1.34 | .34 | 3.35 | .45 | 2.70 | 2.64 | 1.47 |
| Klemp and Wilhelmson (1978) | 2.62 | .65 | 6.55 | 1.22 | 4.28 | 4.18 | 2.62 |
| Orville and Kopp (1977) | 1.34 | ---- | ---- | .50 | 1.64 | ---- | ---- |

Table A2.3 Comparison of evaporation rates in different models in varying conditions. Units of evaporation are g/m³.

These results show that equation (10) produces evaporation rates that compare favorably with those of Tripoli and Cotton (1980) and Orville and Kopp (1977) if drop temperatures are assumed to equal the wet bulb temperature or the values predicted by Beard and Pruppacher (1971). The calculations at $T_{\text{air}} = 277.2$ K and relative humidity = .5 show that (10) predicts values higher than Tripoli and Cotton (1980) by about 20%. The difference is greater at low raindrop water contents and less when the raindrop water content is high. The values predicted by the printed Klemp and Wilhelmson (1978) formula are higher than the others by a factor of about two. This is probably due to confusion resulting from ambiguities in the definition of terms in the published formula.

In summary, a raindrop ventilation coefficient that is linear in raindrop radius is used to develop a simple evaporation formula that produces evaporation rates that agree with the measurements of Beard and Pruppacher (1971) and the modeling formulae of Tripoli and Cotton (1980) and Orville and Kopp (1971).

If the vapor density in the model parcels exceeds the saturation vapor density at the model raindrops' surfaces, condensation onto raindrops occurs, also following (10).
2. Conversion Between Cloud Droplets and Vapor: Evaporation

The evaporation of cloud droplets in model parcels proceeds much more rapidly than the evaporation of raindrops. If a model parcel is sub-saturated and contains cloud water, the cloud water is evaporated until the parcel reaches saturation or until the cloud water is exhausted.

3. Ice Phase Processes: Melting, Warming, Evaporation, Condensation, Sublimation, Deposition

Ice phase precipitation is treated crudely in this model. Ice particles are assumed to be composed of ice and cold water at 273.2 K. The initial mix of solid and liquid phase water can be specified for each downdraft. The temperature assumption is valid for the downdrafts observed in this study, but would have to be modified in order to accurately model downdrafts that extend above the melting level or very deep hailshafts extending to the ground. A fraction of the ice mass is converted to cold water mass (at 273.2 K) during each model timestep. The rate is proportional to the difference between the wet bulb temperature and 273.2 K, with a variable proportionality constant. This process attempts to simulate the melting of ice particles and can be adjusted so that the modeled ice mass melts in a layer that corresponds to an observed melting layer thickness.

The cold water at 273.2 K is converted to normal raindrop water in an attempt to simulate the warming of completely melted ice particles.

If water at 273.2 K is present, evaporation or condensation proceeds according to the raindrop evaporation formula (10) except $\rho_{sv}$ is calculated assuming a drop temperature of 273.2 K. If no water at 273.2 is present, then sublimation or deposition proceeds according to (10) except that the latent heat release into the air parcels is adjusted to account for the ice phase transition.

4. Warming of Liquid Water Mass

Cloud droplets and raindrops transfer heat to the air parcels as they warm or cool due to changing air parcel temperatures and relative humidities (see the discussion of raindrop temperatures above).
5. Entrainment

The model assumes that the downdraft is an entraining similarity plume similar to that described in Morton et al. (1956) and Emanuel (1981). The entrainment parameter and initial plume radius can be varied to reproduce the shape of the observed downdraft or can be set according to laboratory-derived values such as those in Morton et al. (1956). The qualities of the entrained air could be specified with an entrainment sounding. Typically either in-cloud or out-of-cloud air was entrained (see Figure R.25 for a simple schematic diagram of the two entrainment regimes). The nature of the entrainment process in actual thunderstorm downdrafts is not well determined, as discussed in Chapter 1. The nature of the entrainment that was observed in the studied downdrafts is discussed in Chapter 3 and Chapter 4, as is the sensitivity of the model to varied assumptions concerning the nature of this entrainment.

If cloudy air is entrained, cloud droplets are entrained with it. The entrained air parcels are assumed to have zero vertical velocity and thus exert a velocity drag on the downdraft that is proportional to the diameter of the downdraft and the quantity of entrained air.

6. Accelerations

The model allows both thermal buoyancy and loading buoyancy to act on air parcels. The thermal buoyancy is calculated relative to a buoyancy sounding that represents the large scale environment. The loading buoyancy is calculated from the total water content in the parcels (raindrops, ice particles, cold water content, and cloud droplets). As noted above, there is a drag associated with entrainment.

A boundary layer is crudely simulated by bringing parcel velocities linearly to zero near the ground.
D. Sensitivity Studies

The sensitivity of the model to variations in downdraft environment and assumed physics is explored. The intent here is not to determine which mechanisms are most likely to produce strong downdrafts by exploring the model parameter space. The motivation is to examine the model's performance and to insure that it is behaving in a reasonable manner to controlled variations in inputs. The sensitivity to variations in poorly known variables such as the entrainment parameter is also tested.

The results of the base case run to which comparisons will be made is shown in Table A2.4.

<table>
<thead>
<tr>
<th>P (mb)</th>
<th>z (km) approx</th>
<th>θ_e (K)</th>
<th>T (K)</th>
<th>RH</th>
<th>w_a (m/s)</th>
<th>B_T (m/s²)</th>
<th>B_L (m/s²)</th>
<th>ρ_v (g/m³)</th>
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<td>4</td>
<td>334</td>
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<td>0</td>
<td>-0.059</td>
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<td>-0.039</td>
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<td>.62</td>
<td>0</td>
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<td>---</td>
<td>18.0</td>
</tr>
</tbody>
</table>

Table A2.4 Model Results: Base Case. In-cloud entrainment. RH is relative humidity, B_T is thermal buoyancy acceleration, B_L is loading buoyancy acceleration.

This simulation closely approximates the observations of cell S and is discussed in Chapter 3 (Table S.3), except that a dry, unmodified boundary layer has been assumed (T_{ground} = 304 K, RH_{ground} = .54). In this base case, the downdraft entrains in-cloud air while above 900 mb, the approximate lifted condensation level from the proximity sounding shown in Figure 3.B.2, and is thus similar to the plumes in Emanuel (1981).
Table A2.5 Model Results: **Wet Sub-Cloud Environment:** in-cloud entrainment. RH is relative humidity, $B_T$ is thermal buoyancy acceleration, $B_L$ is loading buoyancy acceleration. $T_{\text{ground}} = 304$ K, RH$_{\text{ground}} = .54$

Imposition of a wet sub-cloud environment has little effect on the model downdraft, Table A2.5, since it experiences this environment for only the short time that it spends below cloud base (see Figure 5.1 and discussion in Chapter 5). Not surprisingly, the parcels arrive at the ground slightly cooler and wetter.

Table A2.6 Model Results: **Small Entrainment Parameter:** in-cloud entrainment. RH is relative humidity, $B_T$ is thermal buoyancy acceleration, $B_L$ is loading buoyancy acceleration. $\alpha = .15$

An entrainment parameter value of $\alpha=.25$ is used in the simulations presented in Chapter 3 and Chapter 4, consistent with the results of Turner (1986). The model sensitivity to lower values of this parameter is shown in Table A2.6. As would be expected, the parcels arrive at the surface with lower values of $\theta_e$ since they contain more pristine mid-level air. In addition the downdraft is drier than the base case since there is less availability of entrained cloud droplets and raindrop evaporation cannot keep pace with the rapid rate of compressional warming during descent.
The environmental sounding that is used in the studies in Huntsville was launched from the Redstone Arsenal at 1800 Z, about 2.5 hours before the downdraft events were observed. In the base case run and in the simulations in Chapter 3, the lower levels of the sounding have been warmed to account for additional heating that occurred during the afternoon. The correction was 1 K, which was consistent with the surface mesonet observations.

The model is quite sensitive to small changes in environmental conditions, consistent with the results of Proctor (1989). The results of a 1 K reduction in the 800 mb temperature are shown in Table A2.7. The parcels have difficulty descending through the stabilized region near 800 mb. Parcels accelerate downwards and reach velocities of -6.2 m/s at 700 mb before large positive thermal buoyancies slow them to -2.9 m/s just below 800 mb.

Evaporation of raindrops, the effects of which are enhanced due to the deceleration (smaller velocity = more time for evaporation), continues and the combined effects of gradually decreasing buoyancy due to evaporational cooling and continued precipitation loading eventually cause renewed downward acceleration.

Interestingly, the model arrives at the surface with almost identical thermodynamic characteristics and the peak velocities at low levels are almost the same as in the base case. The small velocities at upper levels are clearly distinguishable from those observed in the triple Doppler analyses presented in Chapter 3.
P (mb) & z (km) & $\theta_e$ (K) & T (K) & RH & $w_a$ (m/s) & $B_T$ (m/s²) & $B_L$ (m/s²) & $\rho_v$ (g/m³) \\
650 & 4 & 334 & 280 & .56 & 0 & 0 & -0.59 & 4.3 \\
800 & 2 & 345 & 287.5 & .97 & -12.0 & -0.050 & -0.050 & 12.1 \\
900 & 1 & 348 & 294 & .87 & -14.5 & -0.090 & -0.045 & 15.7 \\
950 & 0.5 & 349 & 298 & .73 & -15.6 & -0.080 & -0.042 & 16.6 \\
1000 & 0 & 350 & 301.5 & .66 & 0 & --- & --- & 18.3 \\

Table A2.8 Model Results: **High Cloud Water Content:** in-cloud entrainment. RH is relative humidity, $B_T$ is thermal buoyancy acceleration, $B_L$ is loading buoyancy acceleration. $\rho_v = 3$ g/m³

Observations in clouds discussed in Chapter 1 show that cloud water concentrations of 1 g/m³ are typical. This value is used in the base case and in the simulations in Chapter 3 and Chapter 4. The model is sensitive to the level of cloud water since cloud drops evaporate quickly and saturate the model parcels at the expense of raindrop evaporation. In a simulation with high cloud water content, Table A2.8, the downdraft is much stronger and wetter than the base case. The velocities are much higher than the observed values presented in Chapter 3. Because of the great quantity of cloud droplets that are entrained, the downdraft is able to maintain saturation during rapid descent. The nearly wet adiabatic descent results in very large downward thermal buoyancy accelerations.

$P (\text{mb})$ & $z (\text{km})$ & $\theta_e$ (K) & T (K) & RH & $w_a$ (m/s) & $B_T$ (m/s²) & $B_L$ (m/s²) & $\rho_v$ (g/m³) \\
650 & 4 & 334 & 280 & .56 & 0 & 0 & -0.118 & 4.3 \\
800 & 2 & 345 & 290 & .77 & 10.6 & .026 & -0.090 & 11.1 \\
900 & 1 & 348 & 296 & .71 & 12.3 & -.020 & -.078 & 14.8 \\
950 & 0.5 & 349.5 & 299.5 & .65 & 13.5 & -.034 & -.073 & 16.4 \\
1000 & 0 & 350 & 301.5 & .65 & 0 & --- & --- & 18.3 \\

Table A2.9 Model Results: **High Raindrop Water Content:** in-cloud entrainment. RH is relative humidity, $B_T$ is thermal buoyancy acceleration, $B_L$ is loading buoyancy acceleration. Reflectivity = 57 dBZ, Raindrop Water Content = 9.2 g/m³

Since the acceleration due to precipitation loading is the largest acceleration term, it is expected that variations in the precipitation water content will have a strong effect on the model downdraft. The effect of increasing the reflectivity from 53 dBZ to 57 dBZ, corresponding to an increase of precipitation water content from 4.8 g/m³ to 9.2 g/m³ (according to the Geotis, 1971 Z-M relation extended slightly beyond its region of validity), is shown in Table A2.9. The downdraft is stronger and despite the increased
downward velocities, the relative humidities in the parcels are nearly the same as the base case due to the increased evaporation from the increased precipitation water content.

<table>
<thead>
<tr>
<th>P (mb)</th>
<th>z (km)</th>
<th>\theta_e (K)</th>
<th>T (K)</th>
<th>RH</th>
<th>w_a (m/s)</th>
<th>B_T (m/s^2)</th>
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Table A2.10 Model Results: **Low Raindrop Water Content: in-cloud entrainment.** RH is relative humidity, B_T is thermal buoyancy acceleration, B_L is loading buoyancy acceleration. Reflectivity = 45 dBZ, Raindrop Water Content = 1.5 g/m^3

When raindrop water content is lowered, the downdraft weakens as shown in Table A2.10 and also in the simulations of cell R in Chapter 3. Despite the lower downward velocity, the downdraft parcels cannot saturate because of the decreased availability of raindrops for evaporation.

<table>
<thead>
<tr>
<th>P (mb)</th>
<th>z (km)</th>
<th>\theta_e (K)</th>
<th>T (K)</th>
<th>RH</th>
<th>w_a (m/s)</th>
<th>B_T (m/s^2)</th>
<th>B_L (m/s^2)</th>
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<td>0</td>
<td>-0.059</td>
<td>4.3</td>
</tr>
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<td>17.7</td>
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Table A2.11 Model Results: **Ice Phase Precipitation in-cloud entrainment.** RH is relative humidity, B_T is thermal buoyancy acceleration, B_L is loading buoyancy acceleration. Reflectivity = 50 dBZ, Ice Phase Water Content = 3.0 g/m^3

Ice phase precipitation produces a slightly stronger modeled downdraft then does liquid phase precipitation, as shown in Table A2.11. The simulated surface air exhibits slightly depressed values of \theta_e compared to the base case run (349 K vs 350 K), but the diluting effects of entrainment minimize the decrease. Water vapor density is slightly lower throughout the depth of the downdraft. Increased vertical velocities and slightly reduced temperatures due to melting decrease the modeled evaporation rates. In addition condensation of water vapor onto cold water occurs in the uppermost regions of the downdraft. The parcel becomes negatively thermally buoyant at a higher level than the base case.
The outflow air from this simulation is almost indistinguishable from the base case, suggesting that surface observations of outflow thermodynamics may not be able to distinguish downdrafts forced by melting of ice from downdrafts forced by other mechanisms. The vertical velocities in this simulation are higher than those observed. A simulation of the downdraft in cell S including ice is presented in Table S.4. Ice is found to be a secondary factor in the downdraft's forcing, contributing less to parcel cooling than raindrop evaporation.

<table>
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<tr>
<th>P (mb)</th>
<th>z (km) approx</th>
<th>$\theta_e$ (K)</th>
<th>T (K)</th>
<th>RH</th>
<th>$w_a$ (m/s)</th>
<th>$B_T$ (m/s$^2$)</th>
<th>$B_L$ (m/s$^2$)</th>
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<td>0</td>
<td>-.059</td>
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<td>1000</td>
<td>0</td>
<td>341.5</td>
<td>302</td>
<td>.50</td>
<td>0</td>
<td>---</td>
<td>---</td>
<td>14.7</td>
</tr>
</tbody>
</table>

Table A2.12 Model Results: Base Case: Out-of-cloud entrainment. RH is relative humidity, $B_T$ is thermal buoyancy acceleration, $B_L$ is loading buoyancy acceleration.

In the base case simulation, the downdraft entrains from inside the cloud. This is consistent with many observations of entrainment in non-precipitating cumuli, as discussed in Chapter 1. There are some observations that show that lateral entrainment from the cloud edges is occurring. In these cases, downdrafts may be mixing with out-of-cloud air during descent. The out-of-cloud air is sub-saturated and provides a potential for further cooling through the evaporation of rain. On the other hand, the out-of-cloud air does not supply the downdraft with easily evaporated cloud drops so the potential for cooling may not be fully realized. The out-of-cloud air has values of $\theta_e$ that are lower than the in-cloud, updraft, values and, therefore, the values of $\theta_e$ in the downdraft are expected to be lower.

Table A2.12 shows a model simulation with out-of-cloud entrainment (compare with the results for the downdraft in cell S in Table S.2). The model downdraft exhibits vertical velocities that are almost identical to those in the base case. This is due to the previously mentioned trade-off between the effects of the entrainment of warm, cloud droplet-laden air and cold, cloud droplet-free air. As expected the descending parcels are much drier than those in the base case and exhibit depressed values of $\theta_e$. 

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APPENDIX 3  Correction of Persistent Saturation at Mesonet Stations

During outflow events many surface mesonet stations observe saturated or nearly saturated conditions. After the events pass, some of the stations observe drying while some of the stations observe saturated conditions for at least twenty minutes. It is believed that the observed persistent saturations are the result of wetting of the humidity sensors in the mesonet stations. Corrections have been applied to the data analyzed in Figure R.7 and Figure S.6 (stations 114 and 115); the errors and the corrections are discussed in this Appendix.

Figure A3.1 shows time series from two mesonet stations that exhibited the persistent saturation behavior and one that did not. Station 23, a PAM type mesonet station (Brock et al., 1986), observed a peak relative humidity of .95 during an outflow event. The observed relative humidity then drops to about .75 during the next 25 minutes. During this time the temperature rises from the outflow minimum of 297 K up to 301 K. $\theta_e$ averages 356 K well before the event. During the period just before and during the outflow, the value drops to 352 - 353 K. After the event has passed, $\theta_e$ immediately rises to 355 K, very near the pre-event level. This station observes conditions that would be expected during a transitory outflow event and is used a model to correct the data at stations which exhibit persistent saturation.

Stations 104 and 111, both Lincoln Laboratory type mesonet stations (Wolfson et al., 1987, Wolfson, 1989), exhibit the persistent saturation effect (A portion of the record of station 104 is shown in Figure R.7). The observed relative humidity jumps to over .90 during the outflows. Unlike station 23, the observed relative humidity remains at over .90 for at least 25 minutes after the outflow has passed. Intriguingly, the relative humidity observed at station 111 drops abruptly from .98 to .80 in just two minutes at the end of the plotted period. The observed humidity remains between .75 and .80 after the plotted period. The relative humidity at station 104 exhibits a similar, abrupt drop several minutes after the analyzed period. The longer term history of relative humidity observed at the stations is shown in the figure insets. The temperature that is observed at these two stations exhibits behavior that is very similar to that observed at station 23. After sharp drops during the outflow (from 304 K to 298 K at station 111 and from 304 K to 298 K (outflow F) then 296 (outflow R)) the temperature slowly rises to 300 K at both stations. This behavior has also been noted in Wolfson (1989). (A comparison of data
from PAM and Lincoln Laboratory mesonet stations is presented in Wolfson and Iacono, 1987.

Since the observed temperatures recover from the outflow minimum values, but the observed relative humidities remain high, \( \theta_e \) appears to rise. Pre-event values of \( \theta_e \) are near 354 K at station 111 and 349 K at both stations. Ten minutes after the end of the event (as determined by the reduction in wind speed, for example), \( \theta_e \) is near 362 K at station 111 and 355 K at station 104. \( \theta_e \) values continue to rise at both stations until the abrupt drying events several minutes later. The figure shows that the value of \( \theta_e \) observed at station 111 drops to 355 K, near the pre-event value, after the drying.

Based on observations of many mesonet records, and consistent with the drying rate that is exhibited by station 23 in Figure A3.1, the data from the Lincoln Laboratory mesonet stations (no PAM stations experienced persistent saturation) were adjusted. An assumed drying rate of .15 per 25 minutes was applied until the relative humidity dropped from the maximum level observed during the outflows to .75. This produced post-event values of \( \theta_e \) that were at or below the pre-event levels. As stated above, the corrections have been applied to the records in Figure R.7 and Figure S.6 (stations 114 and 115). A steeper correction, in which the relative humidity is brought to .75 in just 15 minutes is also shown in Figure S.6 (station 115) in order to illustrate the sensitivity of the record to the assumptions implicit in these adjustments.
APPENDIX 4 Triple Doppler and Dual Doppler Techniques

A. Introduction

This study used direct triple Doppler calculations through the bulk of the depths of the downdraft regions. The radar arrays in Huntsville and, to a lesser extent, in Orlando, provided an unprecedented opportunity to use directly calculated vertical winds through great depths of the atmosphere.

As discussed in Chapter 2, triple Doppler techniques allow the direct calculation of precipitation particle velocities. Triple Doppler results can be extremely accurate in mid-level regions where dual Doppler techniques are most error prone (being far from either ground level or cloud-top level boundary conditions). This can be extremely valuable in diagnosing ice particle mass fluxes through the melting layer and in determining particle velocities in the electrically active regions of thunderstorms above the melting level.

B. Comparison of Direct Triple Doppler and Integrated Dual Doppler Calculations

The data from Huntsville and Orlando provided an opportunity to compare the results of overdetermined dual Doppler and overdetermined triple Doppler calculations. Figure R.13 and Figure R.14 show vertical slices through cell R extending from the ground to 6 km altitude. Figure R.13 shows vertical air velocities, \( w_a \), calculated using the integration of the mass continuity equation upwards from the ground with a boundary condition of \( w_a = 0 \) m/s at the ground. Overdetermined dual Doppler horizontal winds were used in the integration. Figure R.14 shows \( w_a \) calculated by using the direct triple Doppler values of the vertical precipitation velocity, \( w_p \). As discussed in Chapter 2, a precipitation terminal velocity relationship, depending on radar reflectivity and altitude, was used to convert \( w_p \) to \( w_a \).

The difference between the dual Doppler and triple Doppler results are shown in Figure A4.1. The difference field has been squared in order to make all values positive so the magnitude and not the sign of the differences can be compared. The data have been filtered so that regions that exhibited radar reflectivities of less than 20 dBZ have been eliminated. As would be expected from previous discussions, the difference field is large at both low and high altitudes. Squared values of over 25 m²/s², indicating differences of 5 m/s, are common below 3 km altitude and above 5 km altitude. There is a swath between
3 km altitude and 5 km altitude in which the two fields agree to within 2 m/s. The differences below 3 km are due to large errors in the triple Doppler field. Confirmation of this is found in Figure 2.B.14 which shows a larger slice through a super-set of the region shown in Figure A4.1. The displayed squared three-dimensional divergence, discussed in Chapter 2 as a measure of triple Doppler error, is very large below 2.4 km. Above 5 km, where the three-dimensional divergence indicates that the triple Doppler winds are reliable, increasing errors due to integration contaminate the dual Doppler result. In some regions near the melting level, the triple Doppler result is unreliable due to errors in the precipitation terminal velocity relationships that are applied (as discussed in Chapter 2).

Figure F.16 and Figure F.17 show dual Doppler and triple Doppler vertical wind calculations in slices extending from the ground to 4 km through cell F when the cell was in a particularly favorable location that permitted accurate triple Doppler calculations down to below 2 km altitude. The calculation methods that were used to derive the fields are similar to those in Figure R.13 and Figure R.14 described above.

The difference between the dual Doppler and triple Doppler results in this case is also shown in Figure A4.1. As before, the difference field has been squared to provide positive definite values. Below 1 km, the triple Doppler calculation is not accurate and the differences between it and the dual Doppler results are high. Between 2 km and 3.2 km there is a large zone in the highest reflectivity regions of the cell in which the squared differences are generally less than 9 m²/s², indicative of agreement within 3 m/s. Near 4 km altitude, where the dual Doppler calculation is expected to be worst and the triple Doppler calculation the best, the squared differences frequently exceed 25 m²/s², indicating disagreements of more than 5 m/s in the two calculations.

In Orlando, both the triple Doppler and dual Doppler calculations were less accurate because of both the smaller number of radars, and in the particular case analyzed in Chapter 4, the location of the storm on the edge of the radar network. Consequently, the differences in the results of the two methods of calculation show more substantial differences. As described in Chapter 4, substantial filtering was conducted in a zone of inferred poor data quality.

Figure 4.D.4b and Figure 4.D.10 illustrate differences in the triple Doppler and dual Doppler calculations of a vertical wind field in the Orlando downdraft. At the highest regions displayed in these figures (4 km), where the integrated results of dual Doppler are
expected to be poorest, differences of about 6 m/s are typical in the updraft. In Figure 4.D.13 the dual Doppler and triple Doppler results show very large differences, up to 12 m/s, at 4 km altitude. Direct triple Doppler results are not presented at low altitudes; the presented fields use the combined triple-dual calculations discussed in Chapter 2 with the boundary level for the downward integration at 2.6 km and 3.0 km. In the region of this downward integration, the dual Doppler and combined fields agree to within 2 m/s through most of the downdraft. This is not truly an objective test since the boundary level in the Orlando calculations was subjectively determined and one of the factors in its determination was the agreement between the combined and dual Doppler results.

In the presented calculations the dual Doppler results have employed upward integration of the mass continuity equation. More sophisticated schemes where boundary conditions are applied at both the ground and the upper regions of storms have been used, Ray et al. (1980). Integration errors can be reduced with these methods.

C. Errors in Combined Dual Doppler and Triple Doppler Method Using Downward Integration from a Mid-Level or Low-Level Boundary Level

Combined dual Doppler and triple Doppler techniques, which use a boundary condition at the lowest levels where triple Doppler calculations are accurate and then apply integration of the continuity equation downwards, are presented throughout this study and result in the most accurate wind fields. The technique is described in Chapter 2. In these fields, triple Doppler results are used where they are most accurate and integration of mass continuity is used where it results in the smallest errors. In the absence of boundary condition and integration errors, the downward integration would produce zero vertical wind speed at the surface. The deviation from zero in the actual calculations is an excellent indication of the errors that are likely to exist in the entire vertical velocity field. Figure R.15 (corresponding to Figure R.13 and Figure R.14, discussed above) displays a combined vertical velocity field and the displayed values of vertical velocity at the surface range between ±1.5 m/s, indicating that errors of this magnitude may exist in the displayed fields. Typical values are in the ±1 m/s range suggesting that the displayed downdraft vertical velocities are accurate to within 1 m/s. Figure F.16 shows a similar result in cell F. In this case the generally negative residual vertical velocity, suggests that the displayed downdraft (peak velocity -9 m/s) is in error by about -1 m/s to -2 m/s. A variational technique could be used (Ray et al. (1980)) to minimize these errors. Such techniques were not employed here since the inferred errors were quite small and easily determined. The residual surface
level vertical velocities in the fields analyzed in the Orlando cell, shown in Figure 4.D.4b, 4.D.10, and 4.D.13 are generally in the -2 m/s to -3 m/s range. These suggest that the combined dual Doppler - triple Doppler calculation is overestimating downdraft strength by that amount.

D. Summary

Triple Doppler calculations provide a local measurement of three-dimensional wind velocities. The calculations, while susceptible to geometry errors and errors in precipitation particle terminal velocity estimation, are locally determined and not affected by accumulated integration errors. Triple Doppler vertical wind field calculations are best where dual Doppler calculations are worst, far from the boundary conditions involved in dual Doppler integrations. While the downdrafts in this study did not contain significant ice phase precipitation (with the exception of a narrow region in cell S), triple Doppler techniques would be very valuable in determining ice fluxes and forcing due to melting at mid-levels. The combination of dual Doppler and triple Doppler techniques provides the best estimation of the three-dimensional wind fields.
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Figure 1.1 Schematic diagram of four possible downdraft forcing mechanisms. The four panels depict four basic downdraft forcing mechanisms, melting, dry air entrainment followed by evaporation, precipitation loading, and sub-cloud evaporation, and some possible consequences.
Figure 1.2 Relationship between the difference in surface $\theta_e$ and minimum environmental $\theta_e$, and outflow strength in Huntsville. Pre-event surface measurements are from mesonet stations and environmental minimums are from Redstone Arsenal proximity soundings taken an average of two hours before the outflow events. There is almost no relationship with outflow strength. Least squares fits are plotted along with equations and correlation coefficient squared. Spots represent peak outflow gusts and crosses represent maximum differential radial velocities from Clark (1988).
Figure 1.3 \( \theta_e \) profiles for typical and extreme days in Huntsville. On a typical outflow day the minimum \( \theta_e \) observed at the surface is between the surface and minimum levels of \( \theta_e \). On 03 June 1986, the minimum observed value at the surface \( \theta_e \) is at the minimum observed aloft, implicating ice as a cooling mechanism. Thick vertical lines represent minimum surface values of \( \theta_e \).
Figure 1.4  \( \theta_e \) drops in Denver and Huntsville associated with outflows and the difference between the surface and minimum environmental \( \theta_e \)'s. Top: \( \theta_e \) drops in Denver are small, with most clustered near zero. Middle: \( \theta_e \) drops in Huntsville are much larger. About one half of the events have drops larger than the 8 K explainable by cooling due to melting of adiabatic ice contents. Bottom: difference between surface \( \theta_e \) and minimum value of \( \theta_e \) observed aloft in Redstone Arsenal proximity sounding.
Figure 1.5 Relationship between dryness aloft and outflow strength in Huntsville. Dewpoint depressions from the Redstone arsenal proximity sounding at 700 mb and 850 mb show almost no relation to outflow strength. Least squares fits are plotted along with equations and correlation coefficients squared. Spots are dewpoint depressions at 700 mb; crosses are dewpoint depressions at 850 mb.
Figure 1.6  Relationship between surface relative humidity and outflow strength in Huntsville. Measurements of pre-outflow surface relative humidity from mesonet stations show almost no relationship to outflow strength. Least squares fit is plotted along with equation and correlation coefficient squared. Data from Isaminger (pers. comm.).
Figure 1.7 Relationship between surface relative humidity and outflow strength in Denver. Measurements of pre-outflow surface relative humidity from mesonet stations show almost no relationship to outflow strength. Least squares fit is plotted along with equation and correlation coefficient squared. Data from Isaminger (pers. comm.).
Figure 1.8 Relationship between maximum storm radar reflectivity and outflow strength in Huntsville. Measurements of maximum reflectivity show moderate correlation with outflow strength; the strongest outflows were associated with high reflectivity. Least squares fit is plotted along with equation and correlation coefficient squared. Plotted points usually represent several overplotted events with the same reported values of reflectivity and outflow strength. Data from Isaminger (pers. comm.). Reflectivities reported to nearest 5 dBZ.
Figure 1.9 Relationship between maximum storm radar reflectivity and outflow strength in Denver. Measurements of maximum reflectivity show almost no relationship to outflow strength. Least squares fit is plotted along with equation and correlation coefficient squared. Plotted points may represent more than one overplotted event with the same values of reflectivity and outflow strength. Data from Isaminger (pers. comm.). Reflectivities reported to nearest 5 dBZ.
Figure 2.B.1a Five-radar array in Huntsville. Distances between selected radars are indicated.
Figure 2.B.1b Three-radar array in Orlando. Distances between radars are indicated.
Figure 2.B.2 Horizontal and vertical windfield variances for Huntsville radar array. Horizontal slices through 1 km, 2 km, 4 km, 8 km -- left to right; horizontal dimension of full plot area (discernible in top row) is 30 km. Top row is shaded horizontal variances $\sigma^2_u + \sigma^2_v$. Darkest shade at 1 km and 2 km represents values less than 1.5 and at 4 km and 8 km, less than 1.0. Bottom row is contoured vertical variances $\sigma^2_w$, contoured at 2, 4, 6, and 10. Values are all normalized by $1 \text{ m}^2/\text{s}^2$. 
Figure 2.B.3 Individual components of horizontal windfield variances for Huntsville radar array. Horizontal slice near surface with plot area of 30 km x 30 km. Darkest shade represents values less than 0.4. Values are all normalized by 1 m²/s².
Figure 2.B.4 Vertical windfield variance in vertical slice though Huntsville radar array. Vertical slice extends from the ground to 12 km. Horizontal extent is 26 km and dark lines at bottom are 2 km long. Values are all normalized by 1 m²/s².
Figure 2.B.5 Vertical windfield variances for Orlando radar array. Horizontal slices through 1 km, 2 km, 4 km, 8 km; horizontal dimension of full plot area is 30 km x 30 km. Contours at 2, 4, 6, and 10. Values are all normalized by 1 m²/s².
Figure 2.B.6 Various multiple Doppler arrays used in this and previous field programs represented to scale, Huntsville, Orlando smallest. Baseline lengths in km are indicated.
Figure 2.B.7  Vertical windfield variances for Doppler arrays shown in Figure 2.B.6.  Horizontal slices at 2.4 km AGL.  Contours at 2, 4, 6, and 10 units normalized by 1 m²/s².
Figure 2.B.8  Vertical windfield variances for Doppler arrays shown in Figure 2.B.6. Horizontal slices at 9.6 km AGL. Contours at 2, 4, 6, and 10 units normalized by 1 m²/s².
Figure 2.B.9 Actual windfield variances for 2031 Z, 11 July 1986, Huntsville scan. Vertical wind variances normalized to 1 m²/s². Horizontal slice extending for 8 km x 8 km at 2 km altitude. Panel a: actual variances, panel b: theoretical variances assuming no storm motion and five scanning radars, panel c: number of radars that actually scanned through each gridpoint.
Figure 2.B.10 Precipitation Terminal Velocity Relations. Various terminal velocity relationships are shown to agree to within about 1 m/s. The Dyer 10 cm relation applies to hail and is from Dyer (1975). Pasqualucci is from Pasqualucci (1975), Joss + Waldvogel is from Joss and Waldvogel (1970), Rogers is from Rogers (1964).
Figure 2.B.10b Comparison of Precipitation Terminal Velocity Relation with Radar Derived Values. In region between 3 km and 5 km, there is agreement within 2 m/s in Pasqualucci (1975) formula and values calculated by subtracting triple Doppler $w_p$ from dual Doppler integrated $w_a$. Left: terminal velocity, $w_t$ from Doppler data, Center: Pasqualucci (1975) formula, Right: radar - Pasqualucci (1975) terminal velocity. Contours are reflectivity as measured by the CP4 radar.
Figure 2.B.11 Vertical slice through vertical wind velocity variances. Top Right: variance from the upward integration of the mass continuity equation using variance = 0 at ground level. Top Left: variance from downward integration of the mass continuity equation using variance = 4 surface as boundary condition. Bottom: variance from direct triple Doppler. Variances normalized to 1 m$^2$/s$^2$. Values over 10.9 are overflow values and should be interpreted as infinity. Slice extends from the ground to 10 km altitude.
Figure 2.B.12 Reflectivity mass relations. Various Z-M relations are shown. There is close agreement between all the formulas at low Z and between Geotis (1971) and Marshall and Palmer (1948) at all Z. Abbreviated key codes refer to Marshall, et. al (1947) (MLP Rev), and Wexler (1947) (WLP) as discussed in text.
Figure 2.B.13 Predicted and calculated three dimensional divergence. Vertical slice though cell F downdraft at 2031 Z. Panel b shows observed three dimensional divergence squared and smoothed. The number of scanning radars in various regions is indicated. Panel c is the vertical velocity variance field in units normalized to $1 \text{ m}^2/\text{s}^2$. Panel a is the predicted three dimensional divergence based on the vertical velocity variances from panel c. Slice extends from the ground to 6 km altitude and for 8 km horizontally.
Figure 2.B.14 Three dimensional divergence. Vertical slice through cell R downdraft at 2025 Z, extending from the ground to 9 km. Top: squared and smoothed three dimensional divergence. Bottom: number of radars scanning in various regions of the plot.
Figure 2.C.1 Schematic representation of procedure used to load individual radar gates into cartesian grids. Each gate is split into fifteen simulated point measurements that are loaded into cartesian grids into which they penetrate. Multiple hits to individual gridpoints are frequent from individual and adjacent radar gates. The gridded value is calculated as the average of all values that are loaded in the above fashion.
Figure 2.C.2 Filled and unfilled gridded data illustrating inter-tilt gaps in a single gridded field. 10 km x 10 km horizontal slice at 2 km altitude. Panel a: unfilled reflectivity field as measured by the CP3 radar. Panel b: time difference, in minutes, between the time of observations and 2028 Z. Panel c: post-filled reflectivity field.
Figure 2.C.3 Comparison of gridded data from five radars illustrating widely varied scan strategies. Same 10 km x 10 km slice through 2 km altitude as in Figure 2.C.2. Top: post-filled reflectivity field from the CP2, CP3, CP4, FL2, and UND radars. Bottom: time difference, in minutes, between the time of observations and 2028 Z.
3.B.1a Surface chart of southeastern US at 1200 Z 11 July 1986. Station model shows winds with half barb = 5 m/s, top left $\theta_v$, bottom left T, top right $T_d$. The location of the Huntsville radar array is indicated with a spot.
3.B.1b 500 mb chart of southeastern US at 1200 Z 11 July 1986. Station model shows winds with half barb = 5 m/s, top left $\theta_v$, bottom left T, top right $T_d$. The location of the Huntsville radar array is indicated with a spot.
3. B. 2 Sounding at Redstone Arsenal at 1800 Z 11 July 1986. Temperature and dewpoint plots. Windsocks on right. (From Williams et al. (1987))
3.C.1 Surface divergence illustrating the locations of cell R, cell F, and cell S. Plot area is 6 km x 9 km. Surface horizontal divergence is shaded (yellow/orange = divergence, blue = convergence) and reflectivity measured by the CP4 radar at the 3 km level is contoured. The three outflows are labelled.
Figure R.1 Time history of vertical reflectivity structure in cell R. Time history of the maximum vertical extent of 40 dBZ and 50 dBZ from 2013 Z until 2034 Z. Reflectivities are as measured from the CP4 radar.
Figure R.2a  Time history of near surface CP4 reflectivity in cell R. Slices at 600 m (400 m above the ground), horizontal extent 5 km x 5 km, every three minutes from 2016 Z until 2034 Z. Cell becomes visible at 2022 Z near (52,77) and at 2025 Z near (51,80).
Figure R.2a Time history of near surface CP4 reflectivity in cell R. Slices at 600 m (400 m above the ground), horizontal extent 5 km x 5 km, every three minutes from 2016 Z until 2034 Z. Cell is visible at 2028 Z near (49,84) and at 2031 Z near (48,90).
Figure R.2b Time history of mid-level CP4 reflectivity in cell R. Slices at 4000 m, horizontal extent 5 km x 5 km, every three minutes from 2016 Z until 2034 Z. Same horizontal plot areas as Figure R.2a. Cell is visible at 2022 Z near (50,76)
Figure R.2b  Time history of mid-level CP4 reflectivity in cell R. Slices at 4000 m, horizontal extent 5 km x 5 km, every three minutes from 2016 Z until 2034 Z. Same horizontal plot areas as Figure R.2a. Cell is visible at 2028 Z near (46,86).
Figure R.3 Time history of surface divergence in cell R showing peak divergence near (51,75) at 2028 Z. Horizontal surface divergence shaded (yellow/orange = divergence, blue = convergence) with contours of reflectivity for reference. Same plot areas as Figure R.2a. Reflectivity contours as measured by the CP4 radar at 800 m altitude.
Figure R.4: Outflow strength versus time. Differential radial velocity, Delta-V across cell R outflow as measured by the CP4 radar. Error bars represent Delta-V estimation errors and the time range represented by each radial velocity field used in the calculation.
Figure R.5 Cell-relative and ground-relative surface winds at 2028 Z. Surface divergence is contoured for comparison with Figure R.3. Contour values are m/s/gridpoint (200 m) where 1 m/s/gridpoint = .005/s. Same plot areas as Figure R.2a.
Figure R.6  Progress of radar-measured surface outflow across surface mesonet. Latitude and Longitude are labelled on left and bottom axes. Radar gridpoint locations are labelled on right and top axes for the 2028 Z shifted location. Since the radar grids are shifted at 10 m/s to account for storm motion, the mesonet locations shift, relative to the radar grid. The corrected longitudinal locations at various grid times are indicated above the plot. Note that station 104 is located at (52,88) at 2028 Z and at (43,88) at 2031 Z and at (61,88) at 2025 Z.
Figure R.7 Mesonet record at station 104. Twenty minutes of data are plotted. The labelled quantities are: T=temperature, solid line, axes upper left; RH=relative humidity, dashed line, axes upper right; $\theta_e$=equivalent potential temperature, solid line, axes middle left; $\rho_v$=water vapor density, dashed line, axes right middle; $V$=one minute average wind velocity, solid line, axes lower left; $R$=one minute rainfall, shaded area, axis lower right.
Figure R.8 Mesonet record at station 24. Twenty minutes of data are plotted. The labelled quantities are: $T$=temperature, solid line, axes upper left; $\text{RH}$=relative humidity, dashed line, axes upper right; $\theta_e$=equivalent potential temperature, solid line, axes middle left; $\rho_v$=water vapor density, dashed line, axes right middle; $V$=one minute average wind velocity, solid line, axes lower left; $R$=one minute rainfall, shaded area, axis lower right.
Figure R.9 Horizontal slice at 3.2 km showing horizontal wind velocities and horizontal divergence, and vertical velocity at 2022 Z. Downdraft is visible in interior of cell. Plot area is 5 km x 5 km and is the same region as displayed in Figure R.2a for 2022 Z. Contours are CP4-measured reflectivity. Panel a shows shaded horizontal divergence in units of m/s/gridpoint or $1=0.005$ m/s. Vectors show horizontal cell-relative winds entering the cell from the southwest. Directly calculated vertical air velocity is shaded in panel b. Upwardly integrated vertical air velocity is shaded in panel c.
Figure R.10 Horizontal slice at 4.4 km, in the melting layer, showing horizontal wind velocities and horizontal divergence, and vertical velocity at 2022 Z. Strong convergence is visible in interior of cell along with a narrow and weak downdraft in directly calculated field. Plot area is 5 km x 5 km and is the same region as displayed in Figure R.2a for 2022 Z. Contours in each frame are CP4-measured reflectivity. Panel a shows shaded horizontal divergence in units of m/s/gridpoint or $v = 0.005 \, \text{m/s}$. Vectors show horizontal cell-relative winds entering the cell from the south. Directly calculated vertical air velocity is shaded in panel b. Upwardly integrated vertical air velocity is shaded in panel c.
Figure R.11 Horizontal slice at 2.0 km showing horizontal wind velocities and horizontal divergence, and vertical velocity at 2022 Z. Downdraft in interior of cell is visible. Plot area is 5 km x 5 km and is the same region as displayed in Figure R.2a for 2022 Z. Contours in each frame are CP4-measured reflectivity. Panel a shows shaded horizontal divergence in units of m/s/gridpoint or 1=0.005/s. Vectors show horizontal cell-relative winds. Directly calculated vertical air velocity is shaded in panel b. Upwardly integrated vertical air velocity is shaded in panel c.
Figure R.12 Horizontal slice at 1.0 km showing horizontal wind velocities and horizontal divergence, and vertical velocity at 2022 Z. Downdraft is visible in interior of cell in integrated vertical velocity field. Plot area is 5 km x 5 km and is the same region as displayed in Figure R.2a for 2022 Z. Contours in each frame are reflectivity. Panel a shows shaded horizontal divergence in units of m/s/gridpoint or 1=0.005/s. Vectors show horizontal cell-relative winds. Directly calculated vertical air velocity is shaded in panel b. Upwardly integrated vertical air velocity is shaded in panel c.
Figure R.13 and Figure R.14. North-south vertical slices through downdraft extending from the ground to 6 km. R.13: left: Vertical velocity calculated directly using triple Doppler is shaded. Downward motion exists near 4 km. R.14: right: Vertical velocity calculated by integrating mass continuity from the ground is shaded. Downdraft extends from 4 km to ground in region of highest reflectivity. Contours are reflectivity as measured by the CP4 radar.
Figure R.15 and Figure R.16. North-south vertical slice through downdraft extending from the ground to 6 km. R.15: left: Vertical velocity calculated using direct triple Doppler above 3.4 km and integrating mass continuity downwards from 3.4 km is shaded. Downdraft extends from 4 km to ground and is similar to downdraft in Figure R.14. R.16: right: Horizontal divergence is shaded, 1 unit = .005 m/s, with blue indicating convergence and orange indicating divergence. Slice-parallel, cell-relative winds (v,wa) shown with vectors. Convergence and entrainment are visible at 4-6 km. Contours are reflectivity as measured by the CP4 radar.
Figure R.17 Horizontal slice through cell R at 4.4 km altitude, in the melting layer, showing horizontal divergence and horizontal, cell relative, winds. Convergence (blue) in interior of cloud is visible as well as inward flow of air from the south. Slice extends for 5 km x 5 km. Shaded horizontal divergence in units of .005/s. Vectors are cell-relative horizontal velocity and contours are reflectivity as measured by the CP4 radar.
Figure R.18 and Figure R.19. North-south vertical slice through cell R extending from the ground to 6 km. R.18: left: shaded precipitation particle velocities from triple Doppler. Downward velocities are evident below 5 km. Upward motion above 5 km in main cell at (82,28). Contours of reflectivity measured by the CP4 radar. R.19: right: Differential reflectivity field as measured by CP2 radar. High values, indicating liquid, in downdraft below 5 km. Contours of reflectivity measured by the CP2 radar. Slice-parallel precipitation particle wind vectors \((v, w_p)\) are overplotted.
Figure R.20 North-south vertical slice through downdraft extending from the ground to 6 km. Precipitation liquid water content according to Geotis (1971) Z-M relation is shaded. Large differences in precipitation water content arise from the use of different reflectivity observations. Panel a uses CP2 measured reflectivity and panel b uses CP4 measured reflectivity. Reflectivity from the respective radars is contoured.
Figure R.21 North-south vertical slice through downdraft extending from the ground to 6 km. Downward vertical precipitation mass flux is shaded. Panel a uses CP2 measured reflectivity and panel b uses CP4 measured reflectivity to calculate precipitation water content. Reflectivity from the respective radars is contoured.
Figure R.22 North-south vertical slice through downdraft extending from the ground to 6 km. Squared and smoothed three dimensional divergence shaded. Values are small above 2 km altitude, indicating that triple Doppler calculations are reliable. High values, above 10 units (2.5 x 10^{-4} /s^2) delineate region near the ground where triple Doppler calculations are unreliable. Contours are CP4-measured reflectivity.
Figure R.2.3 North-south vertical slice through downdraft at 2028 Z extending from the ground to 6 km. Downdraft still extends from 4 km to the ground in liquid precipitation region. Horizontal convergence is still occurring at 3-5 km altitude. Panel a: horizontal divergence. Panel b: vertical air velocity calculated using direct triple Doppler above 3.0 km and integration of mass continuity below that using the 3.0 km vertical velocity as a boundary condition. Panel c: CP2 measured differential reflectivity. Panel d: vertical air velocity from upward integration of the mass continuity equation. Contours of reflectivity as measured by the CP4 radar.
Figure R.24 North-south vertical slice through the downdraft, extending from the ground to 5 km, at 2031 Z. Downdraft is shallower and weaker and there is very little convergence aloft. Panel a: horizontal divergence. Panel b: vertical air velocity from upwards integration of the mass continuity equation. Panel c: vertical air velocity calculated using direct triple Doppler above 2.0 km and from downwards integration of the mass continuity equation below that using the vertical velocity at 2.0 km as a boundary condition. Contours of reflectivity as measured by the CP4 radar.
Figure R.25 Schematic illustrating different entrainment environments of simulated downdrafts. With outside environment entrainment, air from outside the cloud is mixed with the descending downdraft. With in-cloud entrainment, cloudy air is mixed into the downdraft.
Figure F.1 Time history of vertical reflectivity structure in cell F. Time history of the maximum vertical extent of 40 dBZ, 50 dBZ and 55 dBZ from 2013 Z until 2034 Z. Reflectivities are as measured by the CP4 radar.
Figure F.2 Time history of surface divergence under cell F. Horizontal surface divergence shaded (yellow/orange = divergence, blue=convergence) with contours of CP4 measured reflectivity for reference. Growth of divergence under cell F can be seen near (60,90) from 2025 Z - 2034 Z.
Figure F.3 Cell F outflow strength versus time. Differential radial velocity, $\Delta V$, across cell F outflow as measured by CP4 radar. The outflow strengthens rapidly after 2022 Z. Error bars represent $\Delta V$ estimation errors and the time range represented by each gridded radial velocity field.
Figure F.4 Progress of radar-measured surface outflow across surface mesonet. Latitude and Longitude are labelled on left and bottom axes. Radar gridpoint locations are labelled on right and top axes for the 2028 Z shifted location. Since the radar grids are shifted at 10 m/s to account for storm motion, the mesonet locations shift, relative to the radar grid. The corrected longitudinal locations at various grid times are indicated above the plot. Note that station 104 is located at (52,88) at 2028 Z and at (43,88) at 2031 Z and at (61,88) at 2025 Z.
Figure F.5 and F.6 North-south vertical slices through cell F at 2022 Z displaying raw triple Doppler vertical air velocities and number of observing radars. Both slices extend from the ground to 4.0 km, the contoured values are reflectivity as measured by the CP4 radar. Figure F.5: left: vertical air motion calculated directly using triple Doppler methods showing a downdraft below 2 km. Figure F.6: right: number of radars that observed in each region. The three shades correspond to 3, 4, and 5 radars, as labelled, the region sampled five times is visible at lower levels.
Figure F.7 and F.8  North-south vertical slices through cell F at 2022 Z displaying observed squared three dimensional divergence and vertical air velocities calculated by integrating mass continuity upwards from the ground. Both slices extend from the ground to 4.0 km and the contoured values are reflectivity as measured by the CP4 radar. Figure F.7: left: three dimensional divergence, values are high near the surface and near the radar sampling boundary aloft. Figure F.8: right: upwardly integrated vertical winds showing the downdraft confined below 2 km.
Figure F.9 and F.10 North-south vertical slices through cell F at 2022 Z displaying observed horizontal divergence and slice-parallel wind vectors. Both slices extend from the ground to 4.0 km and the contoured values are reflectivity as measured by the CP4 radar. Figure F.9: left: horizontal convergence from 1 km to 4 km and divergence near the surface in the outflow. Figure F.10: right: slice-parallel wind vectors showing downward air motion below 2 km.
Figure F.11 Horizontal slice through the top of the downdraft in cell F at 2022 Z showing horizontal divergence and wind vectors. Left: horizontal divergence in cell F (and cell S, at the bottom edge of the plot area) well away from the reflectivity edge which is off the right edge of the plot. Right: cell-relative horizontal winds. Inflow is visible from right side of plot into the convergence region. There is considerable horizontal vorticity to the west of the convergence region. Contours are CP4-measured reflectivity.
Figure F.12 North-south vertical slices through cell F at 2025 Z displaying horizontal divergence and integrated vertical winds. Left: horizontal divergence, showing surface divergence below 1 km and convergence in the 1-3 km region. Right: vertical air motion calculated by integrating the mass continuity equation upwards from the ground. The downdraft is confined below 2 km. Both slices extend from the ground to 4.0 km and the contoured values are reflectivity as measured by the CP4 radar.
Figure F.13 North-south vertical slices through cell F at 2028 Z displaying horizontal divergence and integrated vertical winds. Left: horizontal divergence, showing surface divergence below 1 km and convergence in the 1-3 km region. Right: vertical air motion calculated by integrating the mass continuity equation upwards from the ground. The downdraft is confined below 2 km. Both slices extend from the ground to 4.0 km and the contoured values are reflectivity as measured by the CP4 radar.
Figure F.14 and F.15 North-south vertical slices though downdraft in cell F at 2031 Z. Figure F.14: left: number of observing radars. Figure F.15: right: squared three dimensional divergence. Three dimensional divergence squared is smaller (at any given altitude) in region sampled by five radars. Both slices extend from the ground to 4.0 km and the contoured values are reflectivity as measured by the CP4 radar.
Figure F.16 North-south vertical slices through the downdraft in cell F at 2031 Z. Left: unsmoothed directly calculated triple Doppler vertical air velocities in regions where $(\mathbf{V} \cdot \mathbf{V})^2 < 4$ units or .0001 m$^2$/s$^2$ showing top of downdraft near 2.5 km altitude. Right: smoothed vertical air velocities calculated by using triple Doppler values above 1.6 km and then using downward integration of the mass continuity equation below that level. Both slices extend from the ground to 4.0 km and the contoured values are reflectivity as measured by the CP4 radar.
Figure F.17 and F.18 North-south vertical slices though downdraft in cell F at 2031 Z. Figure F.17: left: smoothed vertical air velocities obtained from the upward integration of the mass continuity equation. Figure F.18: right: horizontal divergence. The downdraft confined below 2 km and deep convergence region from 1-3 km are visible. Both slices extend from the ground to 4.0 km and the contoured values are reflectivity as measured by the CP4 radar.
Figure F.19 Horizontal slice through the top of the downdraft in cell F at 2031 Z. Left: horizontal divergence in cell F (and cell S, at the bottom edge of the plot area). Right: cell-relative horizontal winds. Inflow is visible at right of plot into the convergence region which is well correlated with maximum reflectivity. Contours are CP4 measured reflectivity.
Figure S.1 Time history of vertical reflectivity structure in cell S from 2013 Z until 2034 Z. Reflectivities are maximum in cell as measured by the CP4 radar. Reflectivity core below 4 km drops during period while reflectivity aloft is still growing.
Figure S.2a Time history of near surface reflectivity in cell S. Slices at 600 m (400 m above the ground), horizontal extent 5 km x 5 km, every three minutes from 2016 Z until 2034 Z. Cell S can be seen developing near (60,50).
Figure S.2a Time history of near surface reflectivity in cell S. Slices at 600 m (400 m above the ground), horizontal extent 5 km x 5 km, every three minutes from 2016 Z until 2034 Z. Cell S contains highest reflectivity at 2025 Z and 2028 Z.
Figure S.2b Time history of mid-level reflectivity in cell S. Slices at 4000 m, horizontal extent 5 km x 5 km, every three minutes from 2016 Z until 2034 Z. Same horizontal plot area as Figure R. Sa, except for 2031 Z field, which is shifted 5 gridpoints or 1 km to the north. All fields are CP4 measured reflectivity except for 2031 Z measured by FL2 radar and 2034 Z measured by CP3. Cell S develops rapidly near (62,50).
Figure S.2b Time history of mid-level reflectivity in cell S. Slices at 4000 m, horizontal extent 5 km x 5 km, every three minutes from 2016 Z until 2034 Z. Same horizontal plot area as Figure R.Sa, except for 2031 Z field, which is shifted 5 gridpoints or 1 km to the north. All fields are CP4 measured reflectivity except for 2031 Z measured by FL2 radar and 2034 Z measured by CP3. Cell S maintains high reflectivity throughout period.
Figure S.3  Time history of surface divergence in cell S. Strong divergence develops after 2025 Z near (65,80). Horizontal surface divergence shaded (yellow/orange = divergence, blue = convergence) with contours of reflectivity for reference. Divergence values are in m/s/gridpoint (200 m) where 1 m/s/gridpoint = .005 /s Same plot areas as Figure S.2a (note that the plot area shifts north by 10 gridpoints = 2 km at 2028 Z).
Figure S.4 Outflow strength versus time. Differential radial velocity, ΔV, across outflow as measured by the FL2 radar. The outflow intensifies after 2019 Z and exhibits a ΔV of 24 m/s at the end of the radar study period.
Figure S.5 Progress of radar-measured surface outflow across surface mesonet. Latitude and Longitude are labelled on left and bottom axes. Radar gridpoint locations are labelled on right and top axes for the 2028 Z shifted location. Since the radar grids are shifted at 10 m/s to account for storm motion, the mesonet locations shift, relative to the radar grid. The corrected longitudinal locations at various grid times are indicated above the plot. Note that station 104 is located at (52,88) at 2028 Z and at (43,88) at 2031 Z and at (61,88) at 2025 Z.
Figure S.6 Mesonet record at station 27. Twenty five minutes of data are plotted. The labelled quantities are: T=temperature, solid line, axes upper left; RH=relative humidity, dashed line, axes upper right; $\theta_e$=equivalent potential temperature, solid line, axes middle left; $\rho_v$=water vapor density, dashed line, axes right middle; V=one minute average wind velocity, solid line, axes lower left; R=one minute rainfall, shaded area, axis lower right.
Figure S.6 Mesonet record at station 112. Twenty five minutes of data are plotted. The labelled quantities are: T=temperature, solid line, axes upper left; RH=relative humidity, dashed line, axes upper right; \( \theta_e \)=equivalent potential temperature, solid line, axes middle left; \( \rho_v \)=water vapor density, dashed line, axes right middle; V=one minute average wind velocity, solid line, axes lower left; R=one minute rainfall, shaded area, axis lower right.
Figure S.6 Mesonet record at station 33. Twenty five minutes of data are plotted. The labelled quantities are: T=temperature, solid line, axes upper left; RH=relative humidity, dashed line, axes upper right; $\theta_e$=equivalent potential temperature, solid line, axes middle left; $\rho_v$=water vapor density, dashed line, axes right middle; V=one minute average wind velocity, solid line, axes lower left; R=one minute rainfall, shaded area, axis lower right.
Figure S.6 Mesonet record at station 114. Twenty five minutes of data are plotted. The labelled quantities are: T=temperature, solid line, axes upper left; RH=relative humidity, dashed line, axes upper right; $\theta_e$=equivalent potential temperature, solid line, axes middle left; $\rho_v$=water vapor density, dashed line, axes right middle; $V$=one minute average wind velocity, solid line, axes lower left; $R$=one minute rainfall, shaded area, axis lower right. Multiple lines for RH, $\theta_e$, $\rho_v$ indicate differing assumptions about correct rate of drying after rain ends. This station experienced persistent wetting as discussed in Appendix 3.
Figure S.6 Mesonet record at station 115. Twenty five minutes of data are plotted. The labelled quantities are: T=temperature, solid line, axes upper left; RH=relative humidity, dashed line, axes upper right; θ_e=equivalent potential temperature, solid line, axes middle left; ρ_v=water vapor density, dashed line, axes right middle; V=one minute average wind velocity, solid line, axes lower left; R=one minute rainfall, shaded area, axis lower right. Multiple lines for RH, θ_e, ρ_v indicate differing assumptions about correct rate of drying after rain ends. This station experienced persistent wetting as discussed in Appendix 3.
Figure S.6 Mesonet record at station 37. Twenty five minutes of data are plotted. The labelled quantities are: T=temperature, solid line, axes upper left; RH=relative humidity, dashed line, axes upper right; $\theta_e$=equivalent potential temperature, solid line, axes middle left; $\rho_v$=water vapor density, dashed line, axes right middle; V=one minute average wind velocity, solid line, axes lower left; R=one minute rainfall, shaded area, axis lower right.
Figure S.6 Mesonet record at station 38. Twenty five minutes of data are plotted. The labelled quantities are: T=temperature, solid line, axes upper left; RH=relative humidity, dashed line, axes upper right; $\theta_e$=equivalent potential temperature, solid line, axes middle left; $\rho_v$=water vapor density, dashed line, axes right middle; V=one minute average wind velocity, solid line, axes lower left; R=one minute rainfall, shaded area, axis lower right.
Figure S.7. Maximum one minute winds as measured by surface mesonet. Contours of maximum wind at 7 m/s, 10 m/s, and 13 m/s. Strongest outflow is under deep downdraft of cell S at late stages. Latitude and Longitude are labelled on left and bottom axes. Radar gridpoint locations are labelled on right and top axes for the 2028 Z shifted location. Since the radar grids are shifted at 10 m/s to account for storm motion, the mesonet locations shift, relative to the radar grid. The corrected longitudinal locations at various grid times are indicated above the plot. Note that station 104 is located at (52,88) at 2028 Z and at (43,88) at 2031 Z and at (61,88) at 2025 Z.
Figure S.8 Region of large $\theta_e$ drops during downdraft passage. Contour indicates region with drops of 8 K or more which occurred during later stages of downdraft. Latitude and Longitude are labelled on left and bottom axes. Radar gridpoint locations are labelled on right and top axes for the 2028 Z shifted location. Since the radar grids are shifted at 10 m/s to account for storm motion, the mesonet locations shift, relative to the radar grid. The corrected longitudinal locations at various grid times are indicated above the plot. Note that station 104 is located at (52,88) at 2028 Z and at (43,88) at 2031 Z and at (61,88) at 2025 Z.
Figure S.9 Radar coverage of cell S. The shaded numbers of observing radars shows the poor coverage of cell S at 2025 Z. The southeast of the cell is only sampled by the CP4 radar, while most of the rest is only sampled by three radars. The plot area is 5 km x 5 km. Contours are CP4-measured reflectivity.
Figure S.10. Horizontal slice through lower levels of cell S at 2022 Z near cloud base at 1 km. Weak downdraft is consistent with lack of divergence at ground in Figure S.3. Left: horizontal divergence. Center: vertical air velocity calculated from the upward integration of mass continuity from the ground. Right: vertical air velocity calculated directly using triple Doppler methods, inaccurate at this level. Plot area is 5 km x 5 km. Contours are CP4-measured reflectivity.
Figure S.11. Horizontal slice through lower levels of cell S at 2022 Z at 2 km. Upwards vertical motion indicates downdraft is confined to lowest 2 km. Left: horizontal divergence. Center: vertical air velocity calculated from the upward integration of mass continuity from the ground. Right: vertical air velocity calculated directly using triple Doppler methods. Plot area is 5 km square. Contours are CP4-measured reflectivity.
Figure S.12. Vertical slice through downdraft region in cell S at 2022 Z. Convergence region is in region of upwards motion with possible very weak downdraft below about 1 km. Left: horizontal divergence and slice parallel wind vectors (upwards in convergence region at (55,10)). Center: vertical air velocity calculated from the upward integration of mass continuity from the ground. Right: vertical air velocity calculated directly using triple Doppler methods. Plot extends from ground to 5 km. Contours are CP4-measured reflectivity.
Figure S.13. Vertical slice through downdraft region in cell S at 2022 Z. Five radars sample lower regions while only three sample above 2.5 km so windfield errors are large above 2.5 km. Left: number of sampling radars. Center: smoothed three dimensional divergence (not squared). Right: vertical air velocity calculated directly using triple Doppler methods, unsmoothed and showing overflow condition in three radar region aloft. Plot extends from ground to 5 km. Contours are CP4-measured reflectivity.
Figure S.14. Vertical slice through downdraft region in cell S at 2025 Z. Downdraft confined to levels below 2 km according to integrated and directly retrieved vertical velocities. Left: horizontal divergence showing strong convergence aloft. Center: vertical air velocity calculated from the upward integration of mass continuity from the ground. Right: vertical air velocity calculated directly using triple Doppler methods, not reliable below about 2-3 km. Plot extends from ground to 5 km. Contours are CP4-measured reflectivity.
Figure S.15. Vertical slice through downdraft region in cell S at 2025 Z. Radar coverage is good aloft because of particular three radars that sample, precipitation is liquid phase up to 5 km. Left: number of sampling radars. Center: smoothed three dimensional divergence (not squared). Right: differential reflectivity as measured by CP2 and CP2 measured reflectivity. Plot extends from ground to 5 km. Contours left and center are CP4-measured reflectivity.
Figure S.16. Vertical slice through deep downdraft region in cell S at 2028 Z. Downdraft extends to 3-4 km and convergence is at 2-4.5 km. Left: horizontal divergence and slice parallel wind vectors. Center: vertical air velocity calculated from the upward integration of mass continuity from the ground, core of downdraft extends from $y = 62$ to $y = 65$. Right: vertical air velocity calculated directly using triple Doppler methods, strongest downward vertical velocities are from $y = 61$ to $y = 63$. Plot extends from ground to 5 km. Contours are CP4-measured reflectivity.
Figure S.17. Horizontal slice through downdraft region in cell S at 4 km altitude at 2028 Z showing convergence and apparent engulfment at south edge of cell. Left: horizontal divergence and slice parallel wind vectors (horizontal). Right: number of observing radars indicating that convergence is not by-product of observing boundary. Plot is 5 km square. Contours are CP4-measured reflectivity.
Figure S.18. Vertical slice through downdraft region in cell S at 2028 Z. Downward moving precipitation near melting level is liquid phase. Left: differential reflectivity shaded, total reflectivity as measured by CP2 contoured, only northern region of downdraft is sampled by CP2. Right: vertical particle velocities calculated directly using triple Doppler methods, unsmoothed. Plot extends from ground to 5 km. Contours at right are CP4-measured reflectivity.
Figure S.19. Vertical slice through shallow downdraft region in cell S at 2028 Z. Downdraft extends to 2.5 km and convergence is at 1-3 km. Left: horizontal divergence and slice parallel wind vectors. Center: vertical air velocity calculated from the upward integration of mass continuity from the ground. Right: vertical air velocity calculated directly using triple Doppler methods. Plot extends from ground to 5 km. Contours are CP4-measured reflectivity.
Figure S.20. Vertical slice through deep downdraft region in cell S at 2031 Z. Downdraft extends to 4.5 km and convergence is at 2-4.5 km. Left: horizontal divergence and slice parallel wind vectors, there is little convergence in the downdraft below 2.5 km. Center: vertical air velocity calculated from the upward integration of mass continuity from the ground, core of downdraft extends from y = 67 to y = 70. Right: vertical air velocity calculated directly using triple Doppler methods, strongest downward vertical velocities are from y = 66 to y = 69 in region above 2.5 km where triple Doppler method is reliable. Plot extends from ground to 5 km. Contours are CP4-measured reflectivity.
Figure S.21. Vertical slice through deep downdraft and surrounding regions in cell S extending from the ground to 7 km at 2031 Z. Narrow region of low differential reflectivities, ZDR, suggesting mixed phase precipitation, in downdraft region. Left: ZDR shaded and total reflectivity contoured as measured by CP2, region of low ZDR extends from (68,6) (1.2 km altitude) to (71,20) (4 km altitude), in the region of downward air velocities shown in Figure S.20. Center: unsmoothed vertical particle velocities calculated directly using triple Doppler methods, showing downward precipitation velocities in low ZDR shaft. Right: squared, smoothed, three dimensional divergence, low aloft in downdraft, indicating reliability of directly calculated vertical velocities. Contours at center and right are CP4-measured reflectivity.
Figure S.21b. Ungridded differential reflectivity, ZDR, in downdraft region. Data from individual RHI tilts illustrates the presence of the low ZDR region at the north (left) side of the cell at one time only. Left: total reflectivity, Right: ZDR. At 20:29:31 Z, there is a region of near 1 dB ZDR at 17 km range extending from 1.5 km to 4 km altitude in an area exhibiting about 50 dBZ total reflectivity. At 20:32:24 Z, in nearly the same cell relative location the feature is not visible. Recorded ZDR values are integral, the gridding process creates the fractional values shown in Figure S.21 and elsewhere.
Figure S.22. Vertical slice through shallow downdraft region in cell S at 2031 Z. Downdraft extends to 2 km and horizontal convergence extends from 1-4 km. Left: horizontal divergence. Center: vertical air velocity calculated from the upward integration of mass continuity from the ground. Right: vertical air velocity calculated directly using triple Doppler methods. Plot extends from ground to 5 km. Contours are CP4-measured reflectivity.
Figure S.23. Vertical slice through deep downdraft region in cell S at 2034 Z. Downward precipitation motion near melting level but mostly liquid phase, good radar coverage aloft up to 4 km but poor above that so top of downdraft not well sampled. Left: shaded number of sampling radars. Center: differential reflectivity shaded and total reflectivity contoured as measured by CP2. Right: unsmoothed vertical particle velocities calculated directly using triple Doppler methods. Plot extends from ground to 6 km. Contours are CP4-measured reflectivity.
Figure S.24. Vertical slice through deep downdraft region in cell S at 2034 Z. Downdraft extends to 5 km, past the top of good radar coverage. Left: horizontal divergence and slice parallel wind vectors. Center: vertical air velocity calculated from the upward integration of mass continuity from the ground. Right: vertical air velocity calculated directly using triple Doppler methods. Plot extends from ground to 6 km. Contours are CP4-measured reflectivity.
Figure S.25. Vertical slice through shallow downdraft region in cell S at 2034 Z. Downdraft extends to 2.5 km and convergence is at 1-3 km. Left: horizontal divergence. Center: vertical air velocity calculated from the upward integration of mass continuity from the ground. Right: vertical air velocity calculated directly using triple Doppler methods. Plot extends from ground to 5 km. Contours are CP4-measured reflectivity.
Figure 4.B.1. Surface weather (top) at 1800 Z and 500 mb weather (bottom) at 1200 Z 03 October 1990. Surface pressure contours show easterly flow pattern and observations show humid surface air. Station models show temperature, top left, dewpoint, bottom left, windflag, one barb = 10 knots (5 m/s). At 500 mb there is almost no flow over Florida and conditions are very dry. Station model shows temperature, top left, dewpoint depression, bottom left, x symbolizing relative humidity < .2, windflag, one barb = 10 knots (5 m/s). Study area is marked with a spot.
Figure 4.B.2. Balloon Soundings at nearest stations to Orlando. Soundings taken at Tampa (TBW) and West Palm Beach (PBI) at 1200 Z 03 October 1990, before the downdraft, and 0000 Z 04 October 1990, afterwards. Full barb on wind flags is 10 knots (5 m/s). Temperature and dewpoint plotted.
Figure 4.B.3. National Radar Summary just before downdraft. Radar summary valid at 1935 Z, about 50 minutes before peak outflow strength. VIP levels 1,3,5 contoured.
Figure 4.C.1. Radial data showing early reflectivity development in cell. Top row: ground level PPI's from MIT. Center left: RHI from MIT. Bottom right: ground level PPI from FL2. Bottom left: PPI from MIT. Range rings are every 2 km and every 10 degrees.
Figure 4.C.2. Vertical slice through unfilled gridded data illustrating inter-tilt gaps in reflectivity fields measured by FL2 (left), MIT (center), and UND (right). MIT scanning strategy resulted in small inter-tilt gaps. Slice extends from the ground to 6 km and for 6 km horizontally.
Figure 4.C.3. Time history of vertical development of reflectivity in main cell and in later stage shallow cell. Contours show maximum reflectivity as function of time and height. Solid lines are main cell, dashed lines are late stage shallow cell to the south. Values greater than 55 dBZ shaded. Peak outflow times indicated by arrows at bottom of plot.
Figure 4.C.4. Vertical slices through the MIT-measured reflectivity field during study period. Slices extend from the ground to 8 km. The collapse of the high reflectivity core from 4 km altitude to the ground can be seen.
Figure 4.C.4. Vertical slices through the MIT-measured reflectivity field during study period. Slices extend from the ground to 8 km. The further collapse and weakening of the high reflectivity core can be seen. The development of the later stage southwestern cell is visible in the last time frame at lower levels.
Figure 4.C.5. Time history of MIT-measured reflectivity near the ground. Persistent high reflectivity area is located over outflow (compare with Figure 4.C.7). Plot area 8 km x 8 km.
Figure 4.C.5. Time history of MIT-measured reflectivity near the ground. High reflectivity under outflow at 2028 Z weakens rapidly by 2032 Z. Increased reflectivity at 2036 Z and 2040 Z is associated with shallow southwestern cell. Plot area 8 km x 8 km.
Figure 4.C.6. Time history of MIT-measured reflectivity at 4 km altitude. High reflectivity at early times disappears rapidly as it drops to surface below the level of this horizontal slice (compare with Figure 4.C.5 and Figure 4.C.3). Plot area 8 km x 8 km.
Figure 4.C.6. Time history of reflectivity at 4 km altitude. Reflectivity is low during entire period except for a brief increase at 2036 Z associated with the shallow southwestern cell. Plot area 8 km x 8 km.
Figure 4.C.7. Time history of surface divergence from 2012 Z until 2040 Z. Divergence grows under maximum reflectivity and peaks at 2024 Z - 2028 Z. Only weak divergence evident under southwest cell at 2036 Z and 2040 Z. Plot area 8 km x 5 km. Contours, as in all plots, are reflectivity as measured by the MIT radar.
Figure 4.C.8. Time history of maximum surface divergence and radial velocity difference across outflow as measured by the FL2 radar. Divergence values, open boxes, have been slightly smoothed with a 400 m low pass filter.
Figure 4.C.9. History of surface divergence over Orlando mesonet. Area of surface divergence plotted for different indicated times. Mesonet station, radar and airport locations indicated.
Figure 4.C.10. Mesonet record at station 23. Twenty-five minutes of data are plotted. $\theta_e$ is nearly constant except for the spike that is explained in Appendix 1. The labelled quantities are: $T=$temperature, solid line, axes upper left; $RH=$relative humidity, dashed line, axes upper right; $\theta_e=$equivalent potential temperature, solid line, axes middle left; $V=$one minute average wind velocity, solid line, axes lower left; $R=$one minute rainfall rate, shaded area, axis lower right.
Figure 4.C.11. Mesonet record at station 22. Twenty-five minutes of data are plotted. $\theta_e$ is nearly constant except for the spike that is explained in Appendix 1. The labelled quantities are: T=temperature, solid line, axes upper left; RH=relative humidity, dashed line, axes upper right; $\theta_e$=equivalent potential temperature, solid line, axes middle left; V=one minute average wind velocity, solid line, axes lower left, peak gust velocity is shown with dashed line; R=one minute rainfall, shaded area, axis lower right.
Figure 4.D.1. Time history of horizontal divergence at 3 km altitude from 2012 Z until 2032 Z. Strong convergence is evident in the highest reflectivity from 2012 Z until 2020 Z with divergence to the south at 2012 Z. Plot area 8 km x 6.6 km. Contours are MIT-measured reflectivity.
Figure 4.D.2. Time history of horizontal divergence at 4 km altitude from 2012 Z until 2032 Z. Divergence is evident in highest reflectivity in the updraft at 2012 Z and 2016 Z with convergence to the north. Plot area 8 km x 6.6 km. Contours are MIT-measured reflectivity.
Figure 4.D.3. Time history of horizontal divergence at 2 km altitude from 2012 Z until 2032 Z. Strong convergence is evident at 2012 Z and 2016 Z with weaker convergence at 2020 Z and 2024 Z. Plot area 8 km x 6.6 km. Contours are MIT-measured reflectivity.
Figure 4.D.4. Time history of the vertical extent of the convergence region and its correlation with the descent of the 55 dBZ maximum reflectivity contour. Bold lines indicate top and bottom of major convergence zone and thin lines indicate 55 dBZ contour, from Figure 4.C.3.
Figure 4.D.4b Horizontal divergence and vertical velocity in north-south vertical slice through the cell at 2012 Z. Downdraft is confined below 2 km, air moving into convergence zone and then into updraft or downdraft. Left: horizontal divergence and slice-parallel wind vectors $(v, w_a)$. Center: vertical air velocity calculated by upwards integration of mass continuity from the ground. Right: vertical air velocity calculated by using direct triple Doppler fields above 2.6 km and downwards integration of mass continuity from that level. Contours are MIT-measured reflectivity. Slice extends from the ground to 4 km and 4 km horizontally.
Figure 4.D.5. Number of scanning radars and three dimensional divergence, 2016 Z. Divergence is small in the interior of the cell except for one region with values over 10 units (2.5 x 10^{-4} \text{ m}^2/\text{s}^2). Left: number of scanning radars shaded. Right: squared and smoothed three dimensional divergence. Contours are MIT-measured reflectivity. Plot area is 4 km x 4 km at 3 km altitude.
Figure 4.D.6. Raw vertical particle velocity and air velocity after filtering only. Left: raw vertical particle velocities as calculated from triple Doppler. Right: vertical air velocities calculated from the particle velocities and then deleted in regions where three dimensional divergence is very large, as discussed in Chapter 2. Contours are MIT-measured reflectivity. Plot area is 4 km x 4 km at 3 km altitude.
Figure 4.D.7. Vertical air velocity, horizontal air velocity and horizontal divergence. Air can be seen entering the convergence zone and, in the positive vertical air velocity, rising. Post-filtered and re-filled fields exhibit reasonable looking patterns. Left: filled and smoothed vertical air velocities from triple Doppler analysis. Right: two dimensional divergence smoothed and horizontal wind vectors showing motion into convergence region. Contours are MIT-measured reflectivity. Plot area is 4 km x 4 km at 3 km altitude.
Figure 4.D.8. Vertical winds from triple Doppler and downward integration and horizontal divergence at 2.0 km, 2016 Z. Left: $w_a$ from downwards integration from 3.0 km using triple Doppler boundary condition, downdraft is in interior of cell. Right: horizontal divergence with wind vectors tracing wind entering the cell, primarily from the north. Contours are MIT-measured reflectivity. Plot area is 4 km x 4 km.
Figure 4.D.9. Vertical winds from upwards integration and three dimensional divergence at 2.0 km, 2016 Z. Left: $w_a$ from upwards integration from the ground, showing almost no vertical velocity in the interior of the cell. Right: squared and smoothed three dimensional divergence, showing low values, suggesting reliable triple Doppler calculations, the interior of the cell. Contours are MIT-measured reflectivity. Plot area is 4 km x 4 km.
Figure 4.D.10 Vertical slice showing downdraft as calculated by triple Doppler and downward integration and by upward integration, 2016 Z. The downdraft is confined below about 2.5 km. Agreement in two fields is excellent throughout slice. Left: downwardly integrated vertical winds below 3.0 km level and triple Doppler winds aloft. Right: vertical winds from upward integration. Contours are MIT-measured reflectivity. Plot extends from the ground to 4 km altitude.
Figure 4.D.11  North-south  vertical slice showing unsmoothed reflectivity and horizontal divergence, 2016 Z. Air can be seen entering the high reflectivity regions of the cell and the convergence zone and then descending or ascending. Left: horizontal divergence and slice parallel winds (v,wa) using triple Doppler and downwards integration as in Figure 4.D.8 and Figure 4.D.10(left). Right: unsmoothed MIT-measured reflectivity. Contours are MIT-measured reflectivity (smoothed for contouring). Slice extends from the ground to 4 km altitude and for 4 km horizontally.
Figure 4.D.12 Horizontal divergence and vertical velocity in the cell at 2 km at 2020 Z. Downdraft is evident in both vertical velocity fields. Left: horizontal divergence and horizontal wind vectors. Center: vertical air velocity calculated by upwards integration of mass continuity from the ground. Right: vertical air velocity calculated by using direct triple Doppler fields above 2.6 km and downwards integration of mass continuity from that level. Contours are MIT-measured reflectivity. Plot area 4 km x 4 km.
Figure 4.D.13 Horizontal divergence and vertical velocity in the cell in east-west vertical slice, 2020 Z. Strong convergence region is shown between updraft and downdraft. Left: horizontal divergence and slice-parallel wind vectors (u, wₐ). Center: vertical air velocity calculated by upwards integration of mass continuity from the ground. Right: vertical air velocity calculated by using direct triple Doppler fields above 2.6 km and downwards integration of mass continuity from that level. Contours are MIT-measured reflectivity. Slice extends from the ground to 4 km and 4 km horizontally.
Figure 4.D.14 Horizontal divergence and vertical velocity in the cell in north-south vertical slice, 2020 Z. Air can be seen entering the cell primarily from the north. The parcels travel several km before entering the convergence zone and ascending or descending. Left: horizontal divergence and slice parallel wind vectors \((v, w_a)\). Center: vertical air velocity calculated by upwards integration of mass continuity from the ground. Right: vertical air velocity calculated by using direct triple Doppler fields above 2.6 km and downwards integration of mass continuity from that level. Contours are MIT-measured reflectivity. Slice extends from the ground to 4 km and 4 km horizontally.
Figure 4.D.15 Horizontal divergence and vertical air velocity in the cell at 2 km at 2024 Z. Reasonable agreement in two vertical velocity fields in downdraft region is discussed in text. Left: horizontal divergence and horizontal wind vectors. Center: vertical air velocity calculated by upwards integration of mass continuity from the ground. Right: vertical air velocity calculated by using direct triple Doppler fields above 2.6 km and downwards integration of mass continuity from that level. Plot area 4 km x 4 km.
Figure 4.D.16 Horizontal divergence and vertical velocity in the cell in east-west vertical slice, 2024 Z. Close agreement in two vertical velocity fields exists in main downdraft region which extends up to 2 km. Left: horizontal divergence and slice parallel wind vectors ($u, w_a$). Center: vertical air velocity calculated by upwards integration of mass continuity from the ground. Right: vertical air velocity calculated by using direct triple Doppler fields above 2.6 km and downwards integration of mass continuity from that level. Contours are MIT-measured reflectivity. Slice extends from the ground to 4 km and 4 km horizontally.
Figure 4.D.17 Horizontal divergence and vertical velocity in the cell in north-south vertical slice, 2024 Z. Air is still entering the cell from the north and passing into the collapsing reflectivity core although reflectivities exceeding 45 dBZ only extend as high as 1.6 km. Left: horizontal divergence and slice parallel wind vectors \( (v, w_a) \). Center: vertical air velocity calculated by integrating mass continuity upwards from the ground. Right: vertical air velocity calculated by using direct triple Doppler fields above 2.6 km and integrating mass continuity downwards from that level. Contours are MIT-measured reflectivity. Slice extends from the ground to 4 km and 4 km horizontally.
Figure 4.D.18 Horizontal divergence and vertical velocity in the cell in east-west vertical slice, 2028 Z. While significant disagreements in the two vertical velocity fields exist and are discussed in the text, there are indications of a downdraft extending as high as 2-3 km. Left: horizontal divergence and slice parallel wind vectors (u, w_d). Right: vertical air velocity calculated by upwards integration of mass continuity from the ground. Center: vertical air velocity calculated by using direct triple Doppler fields above 2.6 km and downwards integration of mass continuity from that level. Contours are MIT-measured reflectivity. Slice extends from the ground to 4 km and 4 km horizontally.
Figure 4.D.19 Time of observation in the cell in east-west vertical slice, 2028 Z grid. The radars sample upper regions at different times that are, in the case of UND and particularly MIT, significantly after the times at which lower levels are sampled. MIT samples the 4 km region two minutes after its ground level observation. Left: MIT, Center: FL2, Right: UND. Contours are MIT-measured reflectivity. Times are minutes before or after 2028 Z. Slice extends from the ground to 4 km and 4 km horizontally.
Figure 5.1 Depth of outflow layer during Huntsville events. The altitude at which the differential radial velocity has dropped to 1/2 of the ground level value is defined as the half outflow height. Typical half heights are near 400 m.
Figure A1.1 Top Left: Data from Rossi (1933) show that water vapor pressures near the ground can exceed those at observation height (1-2 m) by 1.5 mmHg (2 mb). Relative humidities can be 6% higher. Top Right: Data from Franssila (1936) show that values of water vapor pressure and relative humidity at 20 cm altitude exceed the values measured at 100 cm by 1 mmHg (1.33 mb) and 5% during the afternoon. Bottom: Data from Brocks (1948) show that the lapse rate of temperature can exceed 100 K per 100 m within one meter of the ground. Figures reproduced from Geiger (1965).
Figure A1.2 \( \theta_e \) jump due to differences in instrumental response times: theoretical response of the Lincoln Laboratory mesonet station temperature and humidity sensors to an instantaneous change in temperature (from 303.2 K to 297.2 K) and relative humidity (from 0.6 to 1.0), mimicking a wet thunderstorm outflow. While the true values of \( \theta_e \) before and after the change are the same, the relative sluggishness of the temperature response causes a transient increase in observed \( \theta_e \). Observed one minute average \( \theta_e \) values are shown in the inset bar charts. The averaging periods are offset from the event time by 0, 10, and 20 seconds (left to right). The observed \( \theta_e \) jump at station 115 is shown in the rightmost inset chart with a line.
Figure A2.1 Schematic representation of processes active in one-dimensional model used in this study.
Figure A2.2 Raindrop ventilation factor as a function of drop radius according to formulations of Beard and Pruppacher (1971), Kinzer and Gunn (1951) and linear formulations proposed in this study.
Figure A2.3 Comparison of evaporation rates as a function of raindrop size using linear ventilation functions and ventilation function from Beard and Pruppacher (1971) for three different raindrop water contents. Dropsize distributions follow Marshall and Palmer (1948). The linear ventilation functions closely approximate the formulation of Beard and Pruppacher (1971).
Figure A2.4 Raindrop temperature and saturation vapor density at the raindrop surface as a function of relative humidity at constant temperature and pressure. Various formulas for raindrop temperature are contrasted with the wet bulb temperature. All formulas predict raindrop temperatures lower than the wet bulb temperature. \( T_{\text{air}} = 293.2 \text{ K} \) and pressure = 1000 mb.
Figure A3.1 Time series of $\theta_e$, relative humidity, and temperature at two stations exhibiting persistent saturation and one that does not. At station 23 the relative humidity drops after the cessation of rain (marked R). At stations 104 and 111, the relative humidity remains over 0.9 for at least twenty minutes after the cessation of rain. Insets show relative humidity during the 2000 Z to 2200 Z period. Both stations exhibit abrupt drying about forty minutes after the event.
Figure A4.1 Differences between integrated dual Doppler and direct triple Doppler vertical windfield calculations. Slices through the cell R downdraft (left), and the cell F downdraft (right) are shown. The difference field has been squared to facilitate comparisons of the magnitude of the differences. In cell R, between 3 km and 5 km there is a broad region in which the two methods of calculation agree to within 4-9 m²/s² (2-3 m/s). Above and below that level the disagreements are much larger. In cell F, there is a region extending from 1 km to 3.5 km altitude where the differences are less than 4-9 m²/s² (2-3 m/s).