An Intense, Quasi-Steady Thunderstorm over Mountainous Terrain.
Part IV: Three-Dimensional Numerical Simulation

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ABSTRACT

A three-dimensional numerical simulation of an intense, quasi-steady left-moving thunderstorm observed over mountainous terrain is presented. The observational analysis of the evolution of convection leading to this storm is presented in Part I, and a detailed analysis of the Doppler radar-observed storm structure is presented in Parts II and III. This storm was particularly interesting because it initially grew in an environment characterized by terrain-induced boundary layer convergence before a massive mesoscale cold front passed underneath. The front cooled and moistened low levels while veering the surface winds to the north, creating a hodograph of winds strongly backing with height. After frontal passage the initial storm cell grew explosively and turned to the left.

The observed storm evolution after the frontal passage was reproduced well by the numerical simulation. An observed secondary updraft, which was not simulated, was attributed to residual effects of the prefrontal environment, which was not considered. The overall success of this simulation led to the conclusion that the storm structure was largely governed by the environmental wind shear and was only weakly influenced by its triggering mechanism.

The microphysical structure was reproduced only moderately well. The model had the greatest difficulty in simulating the echo intensity. This is attributed to the characteristics of the assumed Marshall-Palmer graupel distribution. However, no apparent degrading effects on the dynamical structure were found as a result.

The dynamical structure compared well with that of right-moving cells described observationally and simulated numerically by a number of authors. In particular, it was found that the leftward movement was induced by pressure forces projected to low levels within an anticyclonically rotating updraft in approximate cyclostrophic balance. The rotation was produced by the tilting of horizontal vorticity (associated with the wind shear) into the vertical and subsequent stretching.

Trajectory analysis of updraft and downdraft parcels revealed the existence of both entrainment and pressure-forced downdrafts. It is demonstrated that much of the vertical pressure gradient acceleration of parcels may be accounted for by pressure in approximate hydrostatic equilibrium with the mean density anomaly of the local environment surrounding the parcel.

1. Introduction

At 1700 MDT 19 July 1977, an intense quasi-steady thunderstorm (henceforth referred to as C11) erupted over South Park, Colorado, and behaved curiously different from other storms existing previously on that day. Whereas the other storms lasted for short lifetimes and drifted toward the north, C11 grew very large, became nearly steady and turned toward the west. At the time of C11, participants in the Colorado State University South Park Area Cumulus Experiment (SPACE) observed the region with dual-Doppler and conventional radar (three radars were in operation, although for technical reasons only two were used in the analysis), the NCAR portable automated mesonet (PAM) stations, aircraft, a boundary layer profiler (BLP), and radiosonde. Because of its interesting character, steadiness and proximity to the observation network, C11 was selected for close study by Doppler analysis.

The events leading up to the formation of C11 were studied extensively and reported by Cotton et al. (1982a; hereafter referred to as Part I). In that study it was found that convection began that day in association with a low-level line of convergence where the diurnal thermally driven upslope flow from the east met with the westerly flow associated with the well-mixed boundary layer to the west. Banta (1984) showed that such a convergence zone is a property of the developing daytime boundary layer in that region for most days during the summer. Clouds existing along the convergence zone were found to have lifetimes of about 45 minutes and to drift north-northeastward as the convergence line moved toward the east. New cells formed at regular intervals on the southern end of the convective line. As the southernmost cumulonimbus was reaching maturity, an intense mesoscale front, produced from the outflow of a large thunderstorm complex 100 km to the north, progressed southward through South Park undercutting the surface convergence pattern and causing a veering of the surface winds to a northerly direction at speeds of over 10 m s⁻¹. Coincident with the passage of the mesoscale cold front,
a right- and left-moving pair of convective cells (subsequently labeled C11 and C12) emerged from the southern cumulonimbus system.

The Doppler data taken during the lifecycle of C11 was studied thoroughly by Knupp and Cotton (1982a,b; hereafter referred to as parts 2 and 3). As a result of the Doppler analysis and comparison of observed temperature soundings to PAM surface observations, they proposed the conceptual model of the flow structure of C11 reproduced in Fig. 1. The left-moving character of C11 was hypothesized to occur as a result of C11’s movement toward moist air flowing over the Mosquito Range to the west and because of gust front lifting dominating along the western flank of the storm. It was suggested that northward propagation was prevented by precipitation recycling to the north. A secondary updraft labeled U2 found in the southern portion of the storm was hypothesized to be an updraft associated with the ingestion of the potentially warm air from the west while it was concluded that the main updraft occurred in association with the uplifting of the strong northerly air flow behind the mesoscale cold front.

In order to better understand the observed flows, a three-dimensional numerical model can be employed to simulate many of the observed structures. The flow dynamics may then be studied in an analogy to a laboratory situation. Indeed, a number of investigators have already shown how numerical models can be a valuable tool in understanding real, observed thunderstorm flow fields. The numerical studies of Klemp et al. (1981) and Rotunno and Klemp (1982) quite successfully reproduced the gross features of Doppler radar-observed supercell structure by specifying an initially horizontally homogeneous atmosphere based on a composite sounding and then perturbing it with a strong low-level hot bubble. For low shear situations, Tripoli and Cotton (1980) have shown that the dynamical structure of the simulated cell can be highly dependent upon the way in which a cell is initialized. Indeed, most radar observations of fields of nonsevere cumulonimbus will show many shapes, sizes and intensities existing concurrently. It was concluded that the proper simulation of such cases must involve the specification of what is thought to be unique to the storm of interest. This leads to more complex initialization techniques based on observed and/or hypothesized convergence fields, temperature anomalies or other structures. The studies of Tripoli and Cotton (1980) and Levy and Cotton (1984) are examples of numerical experiments that used dynamic initialization techniques. In all these cases, three-dimensional models

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**Fig. 1.** Conceptual model of the flow patterns in C11 during its intense quasi-steady stage. Streamlines depict airflow (storm-relative) in the given horizontal planes. The arrowed ribbons represent updraft and downdraft circulations. H and L denote regions of strong and weak flow, respectively, at lower (L), middle (m) and upper (u) levels. The hatched region denotes heavy rain (from Knupp and Cotton, 1982a).
have contributed to the understanding of the observed flow fields within the clouds and to the importance of flow fields which may not be observed in the clear air surrounding the cloud.

In this study, the three-dimensional version of the CSU Regional Atmospheric Modeling System (RAMS) will be employed to simulate thunderstorm development and organization in the environment observed after the passage of the mesoscale cold front. The result of this simulation will then be compared to the observed flow fields of C11. It is believed that the interaction between observed storm structures and numerical experiments is a very effective way to gain a solid physical understanding of the observed storm dynamics and the effects of the flow field in which it is embedded on its structure.

In section 2, the cloud model, including the physics, numerical schemes and initialization, will be summarized. In section 3, the results of the post-mesoscale front simulation will be presented. In section 4, predicted dynamic and thermodynamic fields will be compared to direct and indirect observations. Finally, section 5 will summarize the experimental results and draw conclusions concerning the observed structure of C11.

2. The numerical simulation

a. Model physics

The numerical simulation of C11 was carried out using the cloud model version of the CSU RAMS. The model, except for a change to the prediction of perturbation Exner function, similar to Klemp and Wilhelmson (1978a) rather than perturbation dry air density, is similar to an earlier version described by Tripoli and Cotton (1982). Basically, the model uses a quasi-Boussinesq and fully compressible form of the primitive equations cast in two or three dimensions. The thermodynamic terms in the equations of motion are linearized relative to a dry, height-dependent base state. The basic dry physics predict the three velocity components \((u, v, w)\) the perturbation Exner function \((\pi)\) and the potential temperature \((\theta)\). A first-order turbulence closure is used with a buoyancy enhancement applied to the deformation-based eddy viscosity. When moisture is considered, as in cloud applications, ice liquid water potential temperature \((\theta_{li})\) is predicted instead of \(\theta\) and total water mixing ratio \((r_w)\) is also predicted. Potential temperature \((\theta)\), cloud water mixing ratio \((r_c)\) and water vapor mixing ratio \((r_v)\) are then diagnosed from \(r_w\) and \(\theta_{li}\) based on a thermodynamic scheme described by Tripoli and Cotton (1981) in which \(\theta_{li}\) and \(r_w\) are conserved (except for precipitation) and supersaturation is dissalolved. Besides simple cloud-water-only microphysics, three different levels of precipitation microphysics can be used. The first level includes a model for warm rain mixing ratio \((r_w)\) as described by Tripoli and Cotton (1980); the second adds a model for prediction of pristine ice crystal mixing ratio \((r_i)\) and graupel or hail mixing ratio \((r_g)\) (see Cotton et al., 1982b), while the third includes the prediction of aggregated snowflakes \((r_a)\) (see Cotton et al., 1986). Also, as described by Cotton et al., a primary and secondary nucleation model may be used to predict ice crystal concentration. For this study, the full microphysics level-3 option with predicted ice crystal concentration is used.

Although a terrain-following vertical coordinate transformation can be used with this model to simulate the effects of topography, we chose not to do so for the following reasons. First, part 2 suggests that over the relatively flat South Park area, topography was probably not a factor during the time the storm was studied. Second, in order to afford reasonable resolution of the storm over a long enough period of time, it was necessary to perform a Galilean transformation so that the storm did not move quickly relative to the simulation domain.

b. Numerical scheme

The numerical scheme used is the time-split leapfrog scheme described by Tripoli and Cotton (1982). The scheme calculates acoustic and part of the gravity wave fluctuation on a small time step while the advective and diffusion terms are predicted on a large time step. In this study, a modification was made to the scheme described in the earlier paper in which the vertical portion of the acoustic tendency is determined implicitly using the Crank–Nicholson method employed by Durran (1981). This change allows for greater efficiency when the horizontal grid spacing is much larger than that in the vertical. Our tests of this scheme reveal that the implicit treatment of sound waves has no noticeable effect on the solution even when the timestep is reduced by 90%. For this experiment a fourth-order advective scheme was used in all directions. Coriolis accelerations of \(u\) and \(v\) were calculated.

The grid used for the numerical simulation of C11 was a constant 1-km spacing in both horizontal directions and a 0.5-km spacing in the vertical. The simulation domain contained 47 grid boxes in both the meridional \((Y)\) and zonal \((X)\) directions and 34 grid boxes in the vertical \((Z)\) direction. Therefore, the total domain volume was 47 km \(\times\) 47 km \(\times\) 17 km. The grid origin is defined in the center of the horizontal and at the surface in the vertical. For this grid size a long timestep of 6 sec was used in conjunction with a short timestep of 1 sec.

c. Boundary conditions

The lateral boundary condition employed in this experiment was the Orlanski (1976) radiative condition combined with a mesoscale compensation region (MCR)-induced large-scale pressure gradient acceleration as described by Tripoli and Cotton (1982). The MCR is used to simulate the response of the mesoscale to convection within the simulation domain. For this experiment the MCR surrounds the lateral boundaries.
of the computational domain and is one simulation domain length (47 km) across.

For the top boundary condition, $w$, and the vertical divergence, all turbulent fluxes are required to vanish. Since the heating effects to the $\pi$ tendency are dropped in the same manner as Klemp and Wilhelmson (1978a), the associated upper boundary reflection problems described by Tripoli and Cotton (1982) are avoided. To prevent interaction of the convective circulation with this boundary, the top is extended comfortably above the tropopause to 19 km MSL.

At the lower boundary, the surface layer parameterization described by Tripoli and Cotton (1982) was used with the surface heating turned off. This was done because C11 was observed near 1800 MDT when solar heating was unimportant. For frictional effects a roughness length of 5 cm was used.

d. Initial conditions

The predicted quantities are initialized with horizontally homogeneous dynamic and thermodynamic fields based on the adjusted form of the observed South Park post-mesoscale cold front sounding taken at 1710 MDT. This observed and adjusted sounding and hodograph are shown in Figs. 2 and 3.

The Galilean transformation was used to remove a mean meridional wind component of 7 m s$^{-1}$. This was based on observed northward cell movements of that speed prior to the leftward movement of C11.

The initial perturbation used to initiate the simulated storm was based on a scheme described by Thorpe et al. (1982). Since it was postulated in Part II that the primary forcing of C11's updraft was derived from gust front induced uplifting, the simulated storm was initiated by an artificially produced gust front moving with the observed propagation of the early storms. In this study, it is therefore hypothesized that precipitation, produced by the convective complex located in the genesis region of C11 prior to the passage of the mesoscale cold front, cooled the low levels after frontal passage, thus initiating the updraft by forcing the strong northerly flow over the higher density cool air pool. Such an artificial gust front was induced by specifying a cooling function of $-0.02$ K s$^{-1}$ over the lowest 2.5 km in a circular region of radius 3 km positioned as shown in Fig. 4a. The cooling function was maintained until the onset of simulated precipitation at 1500 s integration. The function was then gradually reduced to zero over a period of 300 s as the natural precipitation-induced cooling increased.

3. Results

a. Dynamic evolution

As planned, the initial forcing produced a strong cumulonimbus cloud which lasted a total of 2.5 hours of simulation time. The updraft circulation began within 5 minutes of the initial time and precipitation became intense by 25 minutes simulation time. When the forcing was terminated at 1/2 hour of integration, the storm was self-sustaining. The updraft reached speeds of 30–35 m s$^{-1}$ by 45 minutes of integration and remained quasi steady for the next 1.5 hours. The simulation was terminated at 2.5 hours (9000 s) of integration, at which time considerable weakening of the circulation occurred. In this section, the dynamic evolution of the simulated storm will be presented in some detail.

The circulation field at one hour is displayed in Fig. 4a–f. At 0.8 km AGL, an arc-shaped region of upward motion occurred on the downshear flank of the low level outflow. Note that the region of downdraft has also taken on the arc shape and now lies almost entirely north of the original region of forcing. This indicates that the gust front is now supported by the ongoing storm circulation, rather than possible residual effects of the initialization cooling function. Similar plots of the 0.8 km AGL pressure, thermal, and water loading portions of the vertical acceleration show that the air within the updraft core is negatively buoyant but is accelerating upward due to vertical pressure forces. Relatively low pressure is found within the updraft core ($p'$ is about 2 mb below that of the surrounding region), however, continued pressure lowering above the updraft core supports the vertical acceleration. At this time, equivalent potential temperature analysis shows a lowering of $\theta_e$ at 0.8 km AGL and south of the updraft.
to temperatures found at 3 km AGL (see Fig. 2). This indicates that some air may be descending from this level down to the 0.8 km AGL level.

The airflow at 6.8 km AGL shows that the updraft core contains northerly momentum, while a southerly jet formed laterally east and west of this core. The vertical zonal cross section (Fig. 4e) shows that the southerly momentum appears not to be transported from any level but is in fact accelerating from pressure forces. It was hypothesized in Part II that these jets result when the southerly flow at 6.8 km AGL is diverted laterally around the updraft by pressure forces associated with the updraft. The pressure field displayed in Fig. 4c confirms the existence of some relatively high pressure upwind of the left updraft core.

The pressure analysis at 6.8 km AGL and to a greater extent at 3.8 km AGL (not shown) also show developing low pressure within the left anticyclonic updraft core. Vorticity analysis shows that relative vorticity reaches values of $-7 \times 10^{-3}$ in the anticyclonic core and $9 \times 10^{-3}$ in the cyclonic core at 6.8 km AGL. At 3.8 km AGL, relative vorticities of $-9 \times 10^{-3}$ in the anticyclonic core and $-13 \times 10^{-3}$ in the cyclonic core were predicted. At this level, where the updraft speed is increasing strongly with height, an intense low-pressure anomaly of $-0.75$ mb is found within the anticyclonic core. No such low pressure is found in the cyclonic updraft core to the east. The low pressure within the anticyclonic updraft at 6.8 km AGL is weaker than at lower levels because of the less mature stage of the circulation at upper levels.

Clearly, a storm-splitting process is underway in much the same manner as described by Klemp and Wilhelmson (1978b), Rotunno and Klemp (1982), and Weisman and Klemp (1984). According to their numerical and analytical studies, under the right stability conditions, a strong, vertically backing wind sheaf profile can favor a storm split in which the left-moving storm is dominant. By 1 h simulation, the updraft structure already shows two distinct cells at 6.8 km AGL. The vorticity and pressure structure are also typical of Weisman and Klemp’s simulations producing a storm split, except that the left-moving storm is dominant, rather than the right, as in their studies.

By 5400 s (Fig. 5a-f), two distinct updrafts appear at 0.8 km AGL. The left updraft, as postulated by Weisman and Klemp (1984), has moved considerably southwest of its original position. Since the grid itself is moving northward at 7 m s$^{-1}$ relative to the ground, the left-moving updraft is actually diverting toward the northwest at 7 m s$^{-1}$. The right updraft, on the other hand, has moved about 3 km north of its position at 1 h. This means it is moving northward at about 8.5 m s$^{-1}$ relative to the ground. Pressure analysis at this time shows that a relative low pressure anomaly of 0.25 mb is associated with the left-moving updraft and a low pressure anomaly of about 0.75 mb is associated with the right updraft. At 3.8 km AGL the pressure anomaly is $-0.75$ mb in both updrafts and at 6.8 km AGL the pressure anomaly is $-0.75$ mb in the left updraft and zero in the right. This supports the model of Rotunno and Klemp (1982) which predicts that the left-moving storm should be supported by the vertical pressure gradient while the right-moving updraft will be suppressed by vertical pressure forces. The effect is seen most strongly at 5400 s near 6.8 km AGL, where the left updraft has peak speeds of nearly 30 m s$^{-1}$ while the right updraft attains speeds of only slightly over 20 m s$^{-1}$.

The vorticity pattern at 6.8 km AGL (Fig. 5d) shows continued strong anticyclonic vorticity of $-9 \times 10^{-3}$ s$^{-1}$ in the left updraft. The right updraft still maintains...
Fig. 5. As in Fig. 4 except for 5400 s integration time (initial perturbation not displayed here).
Selected locations of tracers A–G displayed in (a) and (b).
FIG. 6. As in Fig. 4 except for 7200 s integration time (initial perturbation not displayed here). Selected location of tracer H displayed in (a).
Fig. 7. As in Fig. 4 except for 9000 s integration time (initial perturbation not displayed here).
a region of cyclonic vorticity of $9 \times 10^{-3}$ s$^{-1}$, however, a region of anticyclonic vorticity has now appeared on the right portion of the west updraft. A similar pattern is found at 3.8 km AGL. At both 3.8 and 6.8 km AGL the relative vorticity pattern correlates poorly with the eastern updraft core. Perhaps as a result, the eastern updraft is relatively nonsteady and even multicellular at times.

The vertical zonal cross section shows that the flow from levels up to 5 km is being drawn into the circulation of the left updraft. The upper limit of the inflow regime corresponds well with the level of maximum vorticity, seen here by the existence of a southerly jet west of the updraft and a northerly jet east of the center. As expected, the north-south cross section shows air entering the right storm primarily at levels below 2 km. This is because midlevel pressure-induced convergence is weaker or nonexistent in this region. The vorticity in the left-moving updraft can be shown to be sufficient to create a cyclostrophic balance supporting the simulated low pressure. To the right, however, low pressure and vorticity are poorly correlated aloft, consistent with a more cellular and short-lived pressure pattern. The result is that flow is drawn primarily from near the surface where thermally induced convergence can be maintained. A plot of equivalent potential temperature (not shown) shows the lowering of $\theta_e$ south of the right updraft, indicating transport downward from the 3 km AGL level. The circulation of the right-moving updraft can be explained well by the two-dimensional model of Thorpe et al. (1982) which relies on differential vertical momentum transport to provide updraft forcing.

At 7200 s (Fig. 6a-f), the left-moving updraft has remained steady at just over 30 m s$^{-1}$ and moved farther south and west relative to the grid. Relative to the ground, the left updraft has turned more northward and is now moving north-northwest at 6 m s$^{-1}$. The right updraft has weakened considerably near the surface and continued to move northward relative to the grid at 1.5 m s$^{-1}$ or northward at 8.5 m s$^{-1}$ relative to the ground. The cold pool associated with the left-moving updraft has strengthened while that associated with the right-moving updraft has weakened. The downward transport of southerly momentum south of the western cell is leading to a notable kink in the gust front. Apparently, the left-moving system has taken on the characteristics of a left-moving supercell.

At 6.8 km AGL, the anticyclonic vorticity associated with the left-moving updraft has increased to $-11 \times 10^{-3}$ s$^{-1}$ and coincidently located with the peak updraft, which now has weakened slightly to 25 m s$^{-1}$. At the same time, the low perturbation pressure center has strengthened to $-1.0$ mb. This can be expected since an inertial balance is being maintained. The anticyclonic vorticity at 3.8 km AGL, however, has decreased to $-7 \times 10^{-3}$ s$^{-1}$, and the associated low pressure center has weakened slightly and spread laterally. The beginning of a kink in the gust front is the first sign of an impending occlusion. It is suggested that this is brought on by the ingestion of air behind the gust front into the updraft. This weakens the updraft and the associated low pressure to the south, causing it to propagate northward into the warmer $\theta_e$ air which is still capable of being lifted to free convection. In doing so, the updraft tilt increases and the support from above continues to weaken.

Also, at the same time, the right updraft has become more multicellular and much weaker. The updraft speeds are now only 12–15 m s$^{-1}$ at 6.8 km. Its associated vorticity pattern is weak and continues to show little correlation with the updraft center. This may be an indication that the right updraft is supported to some extent by the flow field associated with the dominant left-moving updraft. As a result, no significant pressure perturbation is associated with the center at 6.8 km above the ground.

Vertical cross sections through the left updraft (Fig. 6e-f) clearly show the continued ingestion of air up to 4 km AGL into the updraft core. In the zonal plane, the movement of the peak vortex circulation up to 7 km AGL (from 5 km AGL, $\frac{1}{2}$ h) can be seen. This is all evidence that the updraft is eroding from below.

By 9000 s (2.5 h, see Fig. 7a-f) the occlusion process is strikingly apparent at 0.8 AGL. The primary updraft center at low levels (0.8 km AGL) has weakened and moved west-southwest relative to the grid. Relative to the ground, it has moved north-northwest at 7.5 m s$^{-1}$. The gust front in the vicinity of the left-moving updraft has become grossly kinked and moved north of the supporting dynamic pressure forces aloft. Plots of $\theta_e$ (not shown) depict continued low level cooling southwest of the left-moving updraft in conjunction with subsiding air from the 3 km level forced by rain evaporation and ice melting.

At 3.8 km AGL, the left updraft has weakened to under 8 m s$^{-1}$ and has moved north of the anticyclonic vorticity center. No distinct pressure minimum can be found within the updraft core. At 6.8 km AGL, the left updraft has also weakened considerably, but retains somewhat more vigor. At that level speeds of 15 m s$^{-1}$ remain. The center is still coincident with the peak anticyclonic vorticity center, although there is evidence that the updraft is shifting to the north. As a result, the low perturbation pressure center has weakened to 0.5 mb and appears to be shifting south of the updraft. The vertical cross sections also depict the updraft erosion from below. Because of the obvious demise of the simulated storm, the model integration was terminated at this time.

The time versus height history of the peak updraft and downdraft are given in Figs. 8 and 9. In general, the peak updraft occurs about 9 km AGL and remains steady at 28–33 m s$^{-1}$ until 2 h, when it begins to decay. The first pulse, completed just before 1 h, is the peak development of the original cell. This cell was more severely forced by the specified initial perturbation and
had no other convective motions to interact with. The subsequent weaker pulses were within the left-moving updraft. This updraft regime entrained more midlevel air and had weaker forcing from below, thus weakening its magnitude.

The downdraft evolution clearly shows a low-level response to the updraft pulses aloft, with about a 15-min time delay. Peak downdrafts occurred with the saturation of the initial cell. It is interesting that this downdraft of 9 m s⁻¹ was stronger than the one associated with the initial perturbation at 15 min. The downward motion reaches maximum strength at 0.8 km AGL and originates at about 3 km AGL, which is the level of the θₑ minimum of the unperturbed environment. Interestingly, the strongest downdrafts are aloft at the tropopause level and average 9–10 m s⁻¹, which is somewhat stronger than in the subcloud layer.

b. Trajectory analysis

Air parcel trajectories through the storm are computed by selecting a point in space and time and then integrating, first backward in time to determine parcel origin and then forward to determine the parcel destination. Along the path of the trajectory, a history of eleven quantities is kept. These quantities include the tracer coordinate (x, y, z), the three velocity components (u, v, w), equivalent potential temperature, and the acceleration of w due to virtual temperature perturbation, liquid water loading, ice water loading and vertical θₑ gradient. Following Tripoli and Cotton (1981), equivalent potential temperature is defined as

\[
\theta_e = \theta \left( \frac{r_v L_v}{c_p T_d} - \frac{r_i L_i}{c_p T_d} \right),
\]

along the trajectory where \(\theta_e\) is the equivalent potential temperature, \(\theta\) is the potential temperature, \(r_v\) and \(r_i\) are the vapor and ice mixing ratios, \(L_v\) and \(L_i\) are the latent heats of condensation and melting, \(c_p\) is the specific heat capacity of dry air and \(T_d\) is the dewpoint temperature, which will be close to the saturation temperature. Any variation of \(\theta_e\) along a trajectory will be indicative of mixing or ice precipitation into the parcel.

For trajectory analysis, four points were selected across the left-moving updraft core at 6.7 km AGL and 5400 s time, the positions of which are depicted in Fig. 5b. Figure 5a shows 3 more points selected at the 0.8 km AGL level at 5400 s around the left-moving system. In Fig. 4a, an eighth tracer position is displayed. Although many other trajectory positions were computed, the preceding were selected because they adequately describe the basic flow features found in all trajectories.

In Fig. 10 these trajectories are displayed together. Trajectories A–D depict the origin and destination of parcels across the 6.8 km AGL updraft core. Parcel A, which passed through the east side of the anticyclonically rotating updraft originated in the strong northerly flow of 14 m s⁻¹ flow at 0.6 km AGL. At its apex, trajectory A overshoot the tropopause and finally exited the storm to the south. Parcel B originated to the west of A and slightly higher at 0.9 km AGL. Its initial meridional wind component was −11 m s⁻¹, somewhat weaker than A. Trajectory B remained to the west of A and experienced a greater tendency to accelerate northward. As a result, B exited the domain to the

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**Fig. 8.** Peak updraft contoured as a function of height (Z) and time for first 2 hours of integration time. Contour interval is 2 m s⁻¹.

**Fig. 9.** As in Fig. 8 except for peak downdraft.
southwest after overshooting the tropopause. Figure 11 displays the evolution of $\theta$, $w$ and the vertical acceleration terms along the path of B. It is shown that a strong pressure acceleration initiated the updraft. After the LFC (level of free convection) was attained, pressure acceleration opposed the updraft. When the tropopause was reached, strong negative temperature anomalies accelerated the parcel downward. Evidence of considerable environmental mixing aloft can be seen by a sudden increase in $\theta_v$ as the trajectory passed through the tropopause. This mixing apparently occurred for two reasons. First, the stability is greatly reduced due to the strong adiabatic cooling. Second, the low pressure anticyclonic vortex below becomes a
Fig. 11. Evolution of forcing terms and some parameters along the path of trajectory B. (a) Evolution of indicated terms in vertical equation of motion. (b) Evolution of equivalent potential temperature $\theta_e$, height $z$ and vertical velocity $w$ along path. Times corresponding to symbols in Fig. 10 are shown.

Fig. 12. As in Fig. 11 except for trajectory E.

Fig. 13. As in Fig. 11 except for trajectory F.
high pressure cyclonic vortex aloft, eliminating the effect of inertial stability that was inhibiting mixing. The result was strong mixing in the upper troposphere, which led to a large increase in \( \theta_e \).

Trajectory C, which passed through the weaker southerly flow within the updraft core, originated at 1.8 km AGL to the west-northwest. Having only a weak \(-1.84 \text{ m s}^{-1}\) northerly flow initially, C accelerated toward the north and exited to the northwest of the anvil.

Trajectory D, located in the stronger southerly updraft flow at 6.8 km AGL, originated at 7.3 km AGL to the south. This supports the conclusion drawn in Part II, that the southerly jet occurs from horizontal flow diverting around the updraft core. As the parcel entered the storm, the local vertical pressure gradient forced it to sink at nearly \(-3 \text{ m s}^{-1}\). Strong mixing then increased the \( \theta_e \) of trajectory D by 8 C and led to its eventual entrainment into the updraft core. Because trajectory D only reached upward motions of \(5 \text{ m s}^{-1}\), considerably less tropopause overshoot occurred.

Trajectory E originated east of trajectory A and at the surface. This lowest level air is the cool moist mesoscale outflow from the north. Although its \( \theta_e \) is nearly as warm as that initially found in trajectories A–C, it must pass through a considerable region of negative buoyancy to reach the LFC. Figure 10 shows that E first moved upward and passed through the updraft core at 0.8 km AGL (see Fig. 6a) before falling back downward. Figure 12 displays the vertical accelerations experienced by E. Initially, the pressure gradient force drove E upward. By the time the parcel reached 1.2 km AGL, the downward acceleration due to negative buoyancy exceeded that of the upward-directed pressure acceleration and \( w \) began to decrease. Trajectory E reached a maximum height of 2.5 km AGL where a positive buoyancy relative to the initial environment was achieved. However, because other parcels within the updraft were even warmer, a local hydrostatic equilibrium, characterized by a downward perturbation pressure gradient (or Exner function), and precipitation drag balancing thermal buoyancy were formed. The result was a locally higher LFC for E, which was not attained and which led to the reversal of the updraft motion to downdraft motion. Subsequently, E accelerated downward until 1.5 km AGL was reached. At that level the vertical gradient of perturbation pressure reversed in response to the subcloud-base cooling of the local environment. Although E continued to become more negatively buoyant, the downward motion decelerated. This deceleration occurred in all downdraft trajectories and can be anticipated in other steady convective circuits which have developed a mature pool of cool, high density air beneath the gust front.

Trajectories F, G, and H depict 3 types of entrainment-induced downdrafts which were found to be active. Trajectory F, a rear-flanking downdraft originating at about 2 km AGL, is described quantitatively by Fig. 13. Within F, \( \theta_e \) was initially low at 341 K. Upward acceleration first occurred by the pressure gradient force until the LFC was reached and buoyant acceleration became dominant. Because F entered from the SE, ice precipitation drag increased to oppose buoyant and pressure forces causing F to accelerate downward. The subsequent effects of ice melting and rain evaporation then cooled F substantially between 2.5 km AGL and the surface. Trajectory F reached its highest downward velocity of \(-5.3 \text{ m s}^{-1}\) just above the ground.

Entrainment downdraft G originated at 2.5 km AGL, to the NE of the updraft. Its weak initial southerly drift turned southward by the lowering pressure within the storm. As trajectory G approached the updraft, pressure gradient forces drew the parcel upward, releasing latent heat and warming the parcel. As its buoyancy began to increase, the upward pressure gradient force reversed sign, presumably in response to the influence of the neighboring warm updraft core. At the same time, the precipitation increased the effects of drag. The result was similar to that experienced with E, i.e., trajectory G began to accelerate downward. The downward acceleration persisted despite the still warm (but cooling) environmental relative temperatures it possessed. By 1.5 km AGL, the vertical pressure gradient reversed because of the hydrostatic effects of the mature pool of precipitation-cooled air residing at the surface. At this point, the vertical motion weakened as the parcel continued downward and slowly cooled to ambient temperatures. In essence, the strongest vertical motion achieved by G, resulted from the combined effects of pressure gradient and precipitation drag aloft, while its surface penetration occurred in response to slow cooling by evaporation and melting.

Entrainment downdraft H entered from the southwest at 3.9 km AGL and \( \theta_e = 333 \text{ K} \). Before reaching the storm circulation, its \( \theta_e \) was increased to 339 K by numerically induced vertical mixing.\(^1\) Because H approached the storm from the left, no initial precipitation drag was felt. However, the local vertical pressure gradient in the vicinity of the storm pushed the trajectory downward until it became positively buoyant (relative to the initial environment) and in a local hydrostatic balance. The low pressure within the updraft produced the anticyclonic orbit depicted in Fig. 10. When the parcel finally entered the NE quadrant, the local downward perturbation pressure gradient force was reduced in response to the presence of waterloading. This caused the now warm parcel H to rebound upward to near its original level of 3.9 km AGL. The parcel then left the region of heavy precipitation and entered a region of very warm updraft air. The resulting local vertical pressure force then forced trajectory H downward increasing its buoyancy to over 3 K. As it descended toward the SW, it reentered the graupel (or hail shaft) and began to experience substantial cooling. As with G, H reached a maximum vertical downward

\(^1\) This occurred because of a problem with the eddy viscosity model that does not totally shut down vertical mixing in stable regions.
velocity of 5 m s$^{-1}$ at 1.5 km, but decelerated below 1.5 km AGL because of the upward-directed pressure gradient force caused by the mature pool of cool air residing at the surface.

c. Microphysical evolution

Ice was first nucleated 210 s into the integration. At 516 s of integration, graupel was formed for the first time by heavy riming of ice crystals. Aggregated crystals first formed at 600 s integration and grew in numbers rapidly between 1.2 and 2.2 km AGL. Little growth of aggregates by riming was apparent. By 1400 s, the aggregate population became large enough that riming growth led to the formation of graupel. As the increasing graupel dispersed and settled below the melting layer, the first raindrops appeared at 1422 s integration. Precipitation first occurred as very light graupel but did not become significant until 1680 s in association with the development of a rain shaft.

Throughout the integration, rain was formed primarily by melting graupel and aggregates. Conversion of cloud droplets to form rain was absent in this continental airmass (cloud condensation nuclei were assumed at 600 cm$^{-3}$); however, strong updrafts carried enough melted precipitation aloft to allow significant growth by the collision, coalescence and breakup process below the $-20^\circ$C level. Precipitation reaching the ground consisted primarily of rain, with some light graupel. Predicted surface rainfall amounts on the Galilean transformed grid peaked at 5.8 cm beneath the right-moving storm, which remained nearly stationary relative to the grid. The left-moving cell, which was the most steady and most mobile, dropped as much as 3 cm of precipitation on the moving grid surface.

The development of graupel took place in two regimes. From 0° to $-12^\circ$C, graupel was mostly initiated by conversion from highly rimed aggregated snowflakes. Above the $-12^\circ$C level, graupel developed mostly from highly rimed individual crystals. Graupel growth was through two primary mechanisms. Foremost was growth by riming of cloud water, which was dominant at all levels. Between $-12^\circ$ and $-25^\circ$C about 20% of the graupel growth originated from graupel particles collecting aggregated crystals. Graupel growth by collection of raindrops and ice crystals did occur; however, it was of secondary importance. Below the melting level, graupel melting was significantly enhanced by rain collection and subsequent shedding of rain droplets. As the graupel fell below 1.5 km AGL, nearly as much rain was being intercepted and then shed again by graupel as was being formed by melting.

Aggregate snowflakes, of course, formed entirely from aggregation of individual crystals. There are two primary types of aggregation considered in the model. The first involves the random collision of two ice crystals in which the probability of sticking is by the Hosler-Hallgren (1960) mechanism. This process is strongest near 0°C. The other process occurs in the dendritic production zone between $-12^\circ$ and $-15^\circ$C where dendrite crystals mechanically interact with each other causing aggregation. This process has much higher efficiencies than the first due to the irregular paths that the dendrites take during their fall. Moreover, the higher concentration of ice crystals at the colder temperatures greatly accelerates the aggregation process. In this simulation, the primary production zone for aggregates was indeed between $-12^\circ$ and $-15^\circ$C where this mechanism was operating. At heights above the $-15^\circ$C level, aggregate initiation was negligible. The primary growth of aggregates came from both riming and subsequent collision with other ice crystals. This growth tended to be strongest within the primary production zone between $-12^\circ$ and $-15^\circ$C. This was apparent because this is where the aggregates experienced the highest liquid water contents.

Ice crystals were nucleated by contact nucleation, depositional nucleation and splintering of other ice particles. There were two primary production zones. From 0° to $-12^\circ$C splintering was by far the dominant production mechanism with mean production rates averaging nearly four orders of magnitude higher than all the other mechanisms combined. Between $-12^\circ$ and $-20^\circ$C production of crystals was much weaker and all formation processes were of similar magnitude. It was only in this zone that contact nucleation by Brownian, thermophoretic and diffusiophoretic effects was of a competitive magnitude with the other nucleation and production processes. At temperatures colder than the $-12^\circ$C level, depositional nucleation became increasingly strong. Due to the numerical procedure used, the depositional nucleation was unable to remove all cloud water until nearly $-60^\circ$C in the strong updrafts. This problem has now been corrected by allowing any growth from the vapor phase to have access to all the cloud water phase as well. It should not, however, significantly affect the general dynamics of the simulated cloud.

Within the left-moving cell, there is evidence of graupel recycling into the curved updraft. In Fig. 14, the 1 g kg$^{-1}$ graupel surface is displayed at 2 h simulation time. The graupel surface can be seen to overhang to the north. Because flow is being drawn into the circulation from heights up to 4 km AGL, the forward overhang does not reach the ground. Instead, graupel particles are deposited into the inflow jet and subsequently carried back up into the updraft circulation. Because a constant slope Marshall–Palmer distribution is assumed for the graupel, development of increasingly larger-sized particles cannot take place. Such development in a real storm might take place or, as inferred by Part II in the observed storm, the recycled graupel embryos may lead to beneficial competition for supercooled water resulting in suppression of hail particle growth.

Because of the overall greater intensity of the left-moving cell, higher graupel contents were obtained than in the right-moving cell. The graupel pattern
through the 3.8 km AGL level at 7200 s is displayed in Fig. 15. Peak graupel contents in excess of 5 g kg\(^{-1}\) are found in association with the left cell, while contents of only 3 g kg\(^{-1}\) exist in the right-moving cell. This can be attributed to the stronger support the pressure forces gave to the updraft in the left cell through deeper levels which helped suspend a greater mass of water aloft.

Precipitation cooling of air below the freezing level at 2.5 km was dominated by the effects of graupel melting. Below 1 km AGL, however, rain evaporative cooling exceeded melting-induced cooling.

4. Comparison with observations

The simulated properties of the left-moving updraft cell bear a striking resemblance to those of C11 reported in part II. Observations show that prior to C11, a weak multicellular complex was intercepted by a fast moving mesocold front which undercut the complex as it moved south. This produced the semicircular hodograph of vertically backing winds displayed in Fig. 3. Immediately, the complex intensified rapidly and produced an apparent storm split. The left-moving cell

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**Fig. 14.** Depiction of 1 g kg\(^{-1}\) graupel mixing ratio surface as viewed from the WNW for the South Park steady state storm (3D); time is 5400 s.

**Fig. 15.** Horizontal cross section at 3.8 km AGL showing predicted graupel field. Contour interval is 2 g kg\(^{-1}\).
became C11. Strong intensification of C11 subsequently took place as C11 became quasi-steady moving from 180 deg at 7 m s⁻¹. After 1.5 hours, C11’s reflectivity pattern began to elongate to the west. At this time, C11 turned northward and weakened. Allowing ½ hour for the numerical initialization process, the evolution of the simulated storm closely parallels the observed evolution. In both cases, the storm was born out of an initial split. The leftward movement was similar in both cases; that being toward the west-northwest at about 7 m s⁻¹ and subsequently turning northward after 1.5 hours. The observed intensification of the left-moving cell and weakening of the right-moving cell was well simulated. An observed transition from a somewhat multicellular to a primarily uncellular pattern between 0.5 and 1 hour after initial intensification was also found to some extent in the simulation. Finally, the elongation of the observed echo to the west at the time C11 began to dissipate corresponds to the simulated occlusion process which seemed to be responsible for the weakening.

Based on model predictions of condensate, the simulated radar reflectivity may be diagnosed using the formulae described by Cotton et al. (1983).

The predicted and observed reflectivity patterns at 1, 1.5 and 2 hours of simulation time for 3.8 and 6.8 km AGL and a vertical cross section through the storm are displayed next to observed patterns for 1848, 1915, and 1951 MDT at 4 and 7 km AGL and a vertical cross section shown in Fig. 16. Vertical cross sections demonstrate that the NW overhanging echo above an unbounded weak echo region (WER) is well simulated. The predicted echo size and shape evolution also follows observations reasonably well. In particular, the change from a NE–SW echo axis orientation at 1848 MDT at 4 km AGL to a N–S orientation at 1915 MDT to an E–W orientation at 1951 GMT is reproduced. At 7 km AGL, the observed change from a NW–SE orientation at 1848 MDT to a more N–S orientation at 1915 MDT and then to a NW–SE orientation at 1951 MDT is also reproduced at the 6.8 km AGL simulation level. The observed cell SW of the main cell at 4 km AGL and at 1848 MDT also occurred in the simulation but only at the 30 dBZ level, which is not shown.

One major difference is that the simulated 50 dBZ echo region is considerably deeper than in the observed storm. This is probably due to the simulation of unrealistically large graupel contents aloft. This occurs because of the aforementioned inability of the freezing process to remove all liquid water at temperatures below −20°C in strong updrafts. The remaining liquid water then fuels riming growth of graupel high within the updraft, which in turn affects the reflectivity.

In the vertical, the observed core of the reflectivity maxima tilts downshear (toward the north) with height but is vertically stacked to the south in the simulated echo. This is especially evident at 4 km AGL at 1915 GMT. The assumption of a constant slope Marshall–Palmer graupel size distribution, which does not allow the mean graupel size to vary, may be responsible for this effect. As a result, unnaturally large graupel particles exist in the lower portions of the updraft where graupel is being formed initially. Because the lower updraft tilts upshear, resulting large fall velocities will divert the newly formed graupel toward the south, perhaps prematurely leaving the updraft.

The helical structure of the simulated updraft was observed as well. Comparing Fig. 7b in Part II with Figure 5e, it is found that, in both cases, the lateral southerly jet appears to the left of the updraft which is advecting northerly momentum aloft. The observed relative vorticity and vertical motion patterns at the 7 km AGL level are displayed in Fig. 10b and 10e of Part II. The comparison of this figure with Fig. 5b, d demonstrates the remarkable similarity between the two primary updraft structures.

The simulated peak updraft and downdraft (see Figs. 8 and 9) compare closely with observations. Whereas peak updrafts of 25–30 m s⁻¹ were observed at 7–8 km AGL, updrafts of 25–32 m s⁻¹ were simulated between 7 and 10 km AGL. The observed peak downdrafts of 5–7 m s⁻¹ were also correctly reproduced in the simulation. The simulated surface θ, associated with the downdraft, however, was as much as 5°C too warm, due to the aforementioned numerically induced environmental mixing.

There are, however, some important differences in the secondary updraft structure. Most notable, the simulated storm is unable to reproduce the more transient updraft U2 in C11. This updraft produces the southward extension of the observed echo pattern seen in Fig. 16. This updraft cell was observed to weaken and to move from the west side of the primary updraft to the east side. Because it did not exist within the simulation, it is believed that it originated as part of the multicellular complex prior to the formation of C11 by the major storm split. Other exploratory simulations, not reported here, of cumulonimbus within the environment observed before the passage of the mesoscale cold front, indicated that the observed non-steady storms formed updrafts that moved from the northern to the southern portion of a mesoscale convergence line. As they moved southward, they subsequently decoupled from low-level inflow as precipitation fell on top of the upshear-tilted updraft. In general, the earlier storm cells moved in a northerly direction at about 5 m s⁻¹ relative to the ground. Consider the possibility that C11 and C12 also began this way, however, about the time the precipitation shaft was matur- ing the mesoscale cold front passed underneath. Low-level precipitation cooling then led to new updraft cell initiation similar to that used to initiate a storm in the model. At the same time, the original updraft, which remained to the south of the new cell, decoupled from low levels. The observations of Part II indicated that such decoupling did occur. Also, as C11 turned toward the left, U2 was observed to move relatively
Fig. 16. Predicted and observed radar echoes for three time periods during the storm evolution. The first two columns are for model time 3600 s compared to local radar observations at 1848 MST. The second two columns are for model time 5400 s compared to radar observations at 1915 MST. The last two columns compare model predictions at 7200 s to observations at 1951 MST. Echo contours are drawn at 5 dBZ intervals beginning at 30 dBZ for vertical planes and 40 dBZ for horizontal planes. Echo intensities greater than 50 dBZ are shaded. (Observed echo patterns are taken from Part II.)
right, in the direction of movement of the earlier storms. Therefore, it appears that U2 may have been a remnant of the earlier storm structure left over from the earlier storm complex before the passage of the mesoscale cold front.

5. Summary and conclusions

In conclusion, this study has shown the following. First, the storm structure was largely deterministic and closely governed by the flow and stability properties of the environment. Even though the observed storm experienced a complex initiation scenario, the major features of the storm circulation were reproduced very well using a very simple initialization technique. Second, the storm was dynamically forced. Although Knupp and Cotton (1982a) proposed that low-level precipitation downdrafts and effects of precipitation drag to the north played a key role in causing the storm to move to the left of the mean wind, the simulation results indicate that it was dynamic pressure forcing associated with the rotating updraft that was responsible. In fact, Weisman (personal communication, 1985) has found that splitting and favoring of the left or right updraft (depending upon the hodograph) occurs in a weaker but similar fashion even when the effects of precipitation are totally neglected. Third, the right-moving updraft (called C12 by Part II) was more cellular than the left and weakened by midlevel entrainment and downward forcing by dynamic effects of the sheared environment. As a result, that cell was maintained only by thermally induced surface convergence which produced shorter-lived cells.

The fourth finding of this study is that the vertical acceleration of air parcels brought into the storm circulation are strongly influenced by pressure forces in approximate local hydrostatic equilibrium. The result is that thermodynamic parcel instability cannot be determined relative only to the large scale environment. Instead, local environments are created that may suppress the instability of the less unstable parcels. It is for this reason that the lowest level air never reached the upper storm levels and downdrafts did not accelerate all the way to the surface in the simulation.

Finally, it was found that aggregated snowflakes provided the dominant embryos for graupel formation in the 0° to −20°C layer within the simulated storm, while ice crystals serve as graupel embryos aloft. This does not necessarily suggest that all ice models should necessarily include the aggregation process since the net effect can occur by another route. It does demonstrate, however, that it if it is included, aggregation can have an important influence on ice production.

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