Three-Dimensional Numerical Modeling of Convection Produced by Interacting Thunderstorm Outflows.
Part I: Control Simulation and Low-Level Moisture Variations

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ABSTRACT

The Klemp–Wilhelmson three-dimensional numerical cloud model is used to investigate cloud development along intersecting thunderstorm outflow boundaries. The model initial environment is characterized by a temperature and moisture profile typically found in strong convective situations, and the initial wind field is prescribed by a constant unidirectional shear of 2.9 m s⁻¹ km⁻¹ from 0.8 to 8.9 km, with a constant wind everywhere else. The wind shear vector is perpendicular to the line containing the two initial outflow-producing clouds (which are spaced 16 km apart and are triggered by thermal impulses centered at the top of the boundary layer).

The dynamics of the outflow collision are documented using time-dependent, kinematic air parcel trajectories and thermodynamic data. We find that ambient air in the outflow collision region is literally “squeezed” out of the way as the two outflows collide. Some of this air is lifted to saturation, triggering two convective clouds. The upshear member of the pair has a head start in development, and since the two clouds are growing close together and competing for the same air, the upshear cloud is the strongest. In addition, the downshear cell is suppressed because it grows into the region occupied by the upshear cell’s downdraft and rain region.

By looking at the various terms in the invicid form of the vertical momentum equation, we find that low-level air approaching the gust front along the outflow collision line is forced to rise up and over the cold air pool due to a deflection by the pressure gradient force. A third cloud is triggered along the outflow collision line as a result of this frontal uplifting, which is in contrast to the first two cells which are triggered primarily by the forced uplifting from the outflow collision.

Air parcel trajectories indicate that even though the first two cells along the outflow collision line are triggered by a different mechanism than subsequent cells, the air comprising each updraft core is virtually undiluted, and comes from the same general region (z = 0 ~ 0.3 km). On their way to the cloud updrafts, some low-level air parcels approaching the outflow cross the cold air interface. This is a manifestation of the well-known fact that the gust front is a region of turbulent mixing. Once above the outflow, these air parcels may pass through several updrafts and downdrafts as they traverse the cloud region.

The modeled clouds are found to be sensitive to the low-level (0–1 km) moisture. When the moisture in this layer is increased, the collision line clouds become stronger and the rapidity of new cell development increases markedly. Decreasing the low-level moisture has the opposite effect, to the point that only weak shallow clouds form along the outflow collision line. Furthermore, a decrease in the low-level moisture is accompanied by a decrease in the outflow temperature deficit. This in turn decreases the outflow speed, a result that is consistent with classical invicid density current theory.

1. Introduction

It has long been recognized (e.g., Beebe, 1958; Miller, 1972; Danielsen, 1975; Shapiro, 1982) that synoptic- and mesoscale processes are important in setting the stage for convective storm development (e.g., midtropospheric positive vorticity advection and its associated destabilization of the lower levels, upper-level jet streaks, etc.). However, once convection is present, the organization and movement of the storms are strongly influenced by features in the lowest few kilometers of the atmosphere (e.g., Weaver, 1979). One such feature is the thunderstorm cold air outflow (TO), a pool of evaporatively cooled downdraft air that spreads out along the ground beneath a precipitating cloud. This feature is dynamically similar to submerged density currents (Charba, 1974), and can range in depth from a few hundred meters to several kilometers (e.g., Goff, 1975; Wakimoto, 1982). The typical speed of an outflow’s leading edge, or gust front, is 10 m s⁻¹, although speeds as large as 25 m s⁻¹ have been reported (Goff, 1975).

As an outflow advances along the ground, some of the air approaching it is lifted over the gust front. This lifting process has been identified as one of the primary mechanisms responsible for triggering convective clouds (e.g., Purdom, 1982). The interaction of a gust front with a dryline, a cold front, a sea breeze front, or another gust front is important in determining when and where convective clouds are likely to form (e.g.,

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Purdom, 1976, 1979; Sinclair and Purdom, 1983). Using satellite imagery, Purdom (1982) concluded from a sample of more than 9800 storms that the interaction of a TO with another TO or with other boundaries is a primary means by which convection is initiated. This "convective scale interaction" (Purdom, 1979) involves both the cloud scale (~1–10 km) and the mesoscale (~10–500 km), and is thus often unresolved by the rather coarse network of synoptic scale observing stations (whose average horizontal separation is ~300 km). Consequently, except for a few instances in which convective scale interaction has occurred within special observing networks (as discussed below), details of it are not well understood.

Several observational case studies have confirmed that interacting TOS play crucial roles in determining where deep convection will occur. Holle and Maier (1980) presented a detailed analysis of the events that preceded and followed the collision of two TOS on 15 June 1973 in the Florida Area Cumulus Experiment (FACE) mesonetwork. On this day two vigorous convective systems were present (Fig. 1), one west (A) and one east (B) of the mesonetwork. (The echoes northeast of the network did not appear to play a significant role in the events described.) The precipitation-induced downdrafts from storms A and B produced strong outflows which propagated into the mesonetwork (Fig. 1). The outflows collided at 1510 EDT, triggering a strong convective storm near the center of the mesonetwork. This new cell grew rapidly to 12–15 km in height, and at 1530 EDT produced a weak tornado. Holle and Maier reported that the effects of the seabreeze on this convective cell's formation were insignificant, and that the tornado bearing cloud line "grew rapidly in response to the strong, sustained, and organized surface convergence" [resulting from the outflow collision]. Therefore, in the absence of the outflow interaction, this tornadic storm clearly would not have formed.

In a more recent study, Weaver and Nelson (1982) investigated the deep convection initiated by the collision of TOSs from two nearby supercell thunderstorms in Oklahoma. The storm triggered by the outflow col-

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**Fig. 1.** Reflectivity contours (bold solid lines) of 20, 30, and 40 dBZ at 1505 EDT for the storms west (A) and east (B) of the Florida Area Cumulus Experiment (FACE) mesonetwork on 15 June 1973. Also shown are the isochrones of outflow boundaries across the mesonetwork based on surface wind data. The times of wind shifts at each station are plotted below the station location. Dashed lines represent the outflow from storm A and dotted lines represent the outflow from storm B. The tornado location is marked by the triangle. (Adapted from Holle and Maier, 1980.)
lision produced a short-lived tornado, and as pointed out by the authors, this was the only tornado reported from the entire three-storm complex, despite the fact that both of the initial supercell storms possessed strong mesocyclones. Their results also indicated that storm cell velocities are intimately linked to the movement of the thunderstorm outflow.

Wade and Foote (1982), Fankhauser (1982), and Miller and Fankhauser (1982) discussed multiple outflow interactions from the 22 July 1976 multicell storm over the National Hail Research Experiment (NHRE) data network. Strong low-level convergence (1.0 to 2.0 \(10^{-3}\) s\(^{-1}\)) occurred along the various outflow boundaries, with maximum values (\(\sim 3.0 \times 10^{-3}\) s\(^{-1}\)) located at the outflow intersection points. This enhanced low-level convergence played an integral part in the continued generation of new cells in this convective system.

These observational studies of outflow interactions have been supplemented by simulations from numerical cloud models. Tao and Simpson (1984) found that outflow interaction is a key ingredient in promoting cloud merger in their two-dimensional simulations of GATE (GARP Atlantic Tropical Experiment; GARP stands for the Global Atmospheric Research Program) clouds. Turpeinen and Yau (1981) and Turpeinen (1982) used a three-dimensional cloud model to examine cloud interactions from the same GATE data set. They concluded that the orientation of the interacting clouds to the wind shear vector, as well as the spacing between the clouds, had a significant impact on the behavior of modeled cloud development.

In contrast to these previous modeling studies, which sought to examine cloud behavior on a particular day, this study employs the Klemp and Wilhelmson (1978) three-dimensional cloud model to investigate the more general patterns of cloud development along outflow boundaries which interact in somewhat idealized environments. In this way, we may identify basic features of outflow interactions before proceeding to more realistic (and quite often, more complex) situations.

This paper represents the first in a series which examines several aspects of modeled outflow interactions. The specific questions to be addressed here are

1) How are clouds triggered when two outflows collide (sections 3 and 4)?

2) What dynamics are associated with an outflow collision (section 3)?

3) What are the characteristics of clouds which form along the line of intersection between two outflows (e.g., what are the trajectories of air parcels approaching a gust front, where do the clouds obtain their air, and how do these clouds interact with nearby clouds) (section 4)?

4) What are the effects of low-level moisture changes on clouds triggered by colliding outflows (section 5)?

A description of the cloud model is given in section 2, and a summary of findings and recommendations is presented in section 6.

In Part II of this series (Droegemeier and Wilhelmson, 1985), we discuss the effects of the profile and strength of the vertical wind shear on clouds triggered by colliding outflows. In particular, we examine how cell strength and the time interval between successive cells are related to the environmental wind shear.

Finally, Part III of this series (Wilhelmson and Droegemeier, 1986) deals with the effects of surface friction on outflow behavior and on the characteristics of clouds triggered by colliding outflows.

2. Model description and experiment methodology

The model used in these experiments is the Klemp and Wilhelmson (1978) three-dimensional numerical cloud model. We employ the modifications to the grid structure and finite difference equations reported by Wilhelmson and Chen (1982) and used by Droegemeier (1982). In the simulations presented here, a vertically stretched grid is used in which the model vertical resolution varies smoothly from 0.1 km in the boundary layer to \(\sim 1\) km at the top of the domain. This allows the 0.6–0.9 km deep outflows generated by the model in these experiments to be represented with approximately six gridpoints in the vertical, and at the same time gives adequate resolution in the regions of cloud development above. The grid size is 1.0 km in each horizontal direction. Table 1 lists the values of some of the physical and computational parameters used in the experiments. (The symbols are the same as in Klemp and Wilhelmson, 1978.)

<table>
<thead>
<tr>
<th>Parameter</th>
<th>Symbol</th>
<th>Value</th>
</tr>
</thead>
<tbody>
<tr>
<td>Large time step</td>
<td>(\Delta t)</td>
<td>10.0 s</td>
</tr>
<tr>
<td>Small time step</td>
<td>(\Delta t)</td>
<td>3.3 s</td>
</tr>
<tr>
<td>Coriolis parameter</td>
<td>(f)</td>
<td>0.0 s(^{-1})</td>
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<td>Intrinsic gravity wave phase speed</td>
<td>(C_0)</td>
<td>150 m s(^{-1})</td>
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<tr>
<td>Nondimensional surface drag coefficient</td>
<td>(C_d)</td>
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</tr>
<tr>
<td>Nondimensional constant</td>
<td>(C_m)</td>
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<tr>
<td>Dissipation coefficient</td>
<td>(C_r)</td>
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</tr>
<tr>
<td>Nondimensional constant</td>
<td>(C_k)</td>
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</tr>
<tr>
<td>Eddy momentum mixing coefficient</td>
<td>(K_h)</td>
<td>computed</td>
</tr>
<tr>
<td>Eddy mixing coefficient for scalars</td>
<td>(K_m)</td>
<td>(K_d/K_m)</td>
</tr>
<tr>
<td>Turbulent Prandtl number</td>
<td>(K_d/K_m)</td>
<td>1.0</td>
</tr>
<tr>
<td>Fourth-order horizontal mixing coefficient</td>
<td>(K_4)</td>
<td>(1.0 \times 10^9) m(^4) s(^{-1})</td>
</tr>
<tr>
<td>Second-order vertical mixing coefficient</td>
<td>(K_2)</td>
<td>(100–1000) m(^2) s(^{-1})</td>
</tr>
<tr>
<td>Rainwater autoconversion threshold</td>
<td>(a)</td>
<td>3.0 g kg(^{-1})</td>
</tr>
<tr>
<td>Initial thermal impulse</td>
<td>(\Delta \theta)</td>
<td>2.00 K</td>
</tr>
<tr>
<td>Magnitude</td>
<td></td>
<td></td>
</tr>
<tr>
<td>Horizontal radius</td>
<td>(X_r, Y_r)</td>
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<tr>
<td>Vertical radius</td>
<td>(Z_r)</td>
<td>0.50 km</td>
</tr>
<tr>
<td>Height of center above the surface</td>
<td>(Z_c)</td>
<td>1.00 km</td>
</tr>
</tbody>
</table>
The vertical wind shear in these experiments is unidirectional and is oriented perpendicular to the line joining the two initial clouds. (The limitations of these assumptions are addressed below.) Since the Coriolis parameter is set to zero, we make use of the symmetry in the model to reduce the computational domain to one-half the size of the physical domain (Fig. 2). The physical domain (shaded plus unshaded) is 60 km x 50 km in the east–west and north–south directions, respectively. For clarity in discussion, compass directions have been assigned; however, these have no physical basis in a model with a unidirectional wind shear since any orientation of the associated straight line on a hodograph is possible. The model domain is moved at a constant velocity sufficient to keep the convection of interest as far away as possible from the lateral boundaries. The sensitivity of cloud development to the proximity of an open lateral boundary is discussed in Part II of this series.

In all simulations, the initial cloud in the model is triggered by a symmetric thermal perturbation located at the top of the boundary layer (see Table 1 for details), and placed 8 km from the symmetry plane (Fig. 2). Because of the symmetry about this plane, a mirror image thermal is also present as shown in the figure, and thus the outflows from the two resulting initial storms collide exactly along this symmetry plane, hereafter referred to as the outflow collision line (CL). The 16 km separation between the two initial clouds is not only economical, but it is also large enough that the life cycles of these two clouds will not affect each other. (Note that the CL is not an open lateral boundary.)

Although simulating the interaction of two perfectly identical clouds is physically unrealistic, it does provide an initial basis for examining the questions listed in section 1. We felt that the most natural way to proceed with these experiments was to start with the simplest setting, and thus gain an understanding of the basic features present in outflow interactions. Therefore, the geometrical configuration of these experiments was deliberately made simple to satisfy this goal, and care must be exercised in generalizing these results to more complex settings. Finally, the orientation chosen for the shear vector allowed us to reduce the size of the computational domain. Initial experiments were performed using both perpendicular and parallel orientations, but economic limitations suggested that a more complete set of experiments could be conducted using only one orientation. In the future, these relatively simple conditions can be relaxed (e.g., by varying the initiation time and intensities of the two initial thermals, by changing their orientation to the vertical shear vector, or by using two-dimensional wind profiles).

The sounding used to initialize the model is shown on a Skew-T–logP diagram in Fig. 3. (All model initial fields are horizontally invariant except for the initial thermal perturbation.) This sounding represents a synthesis of several inland Florida soundings taken on days defined as being “unsettled” (Watson et al., 1981; Ron Holle, personal communication, 1981). Outflow interactions are known to occur in Florida under such conditions (Holle and Maier, 1980). Figure 3 shows that the environmental potential temperature is dry adiabatic (θ = 303 K) from the surface to 1.1 km, and from there up to 13.7 km, the profile is nearly moist adiabatic. The mixing ratio qe is constant (19 g kg⁻¹) from the surface to 0.7 km, at which point it decreases with height. The lifted condensation level, convective condensation level, and level of free convection all occur at 0.9 km, and the equilibrium level of a parcel lifted from the surface is ~14.5 km with a θe of ~356 K (note that a parcel at the top of the constant qe layer needs to be lifted, neglecting entrainment, only ~0.25 km to reach saturation). The structure of this sounding, which has a lifted index of ~ -6 K, is indeed well-suited for strong convective activity.

In the early stages of this study, we planned to initialize the model with Florida-type wind profiles taken from days when outflow interactions were known to have occurred. However we abandoned this philosophy and, as mentioned earlier, decided to take a more general approach. The unidirectional wind shear profile used in this study, shown in Fig. 4, is an idealized profile whose low-level shear resembles a Florida environment, and whose upper-level shear is much stronger than typically found in Florida. The wind is constant below cloud base (~0.8 km), and increases linearly with height (constant shear of -2.9 m s⁻¹ km⁻¹) to 8.9 km (note that the downshear direction is toward the south). In Part II of this series, we examine several wind shear profiles and shear strengths. However, we will confine our discussion here to the shear profile of Fig. 4, and use it as the control shear. (In keeping with the convention adopted in Part II, the control simulation is
in cloud models (e.g., Weisman and Klemp, 1982), results from the fact that the cloud is growing in an unstable environment devoid of other clouds. The onset of precipitation occurs very quickly in the initial cloud, and drag effects induced by precipitation loading in the updraft cause a downdraft ($w_{\text{min}} \sim -13$ m s$^{-1}$) to form immediately south (downshear) of the updraft. The sensitivity of a modeled cloud’s downdraft and associated outflow in this type of environment to simple changes in the precipitation parameterization are addressed by Droegemeier (1982). He found that the downdraft of the model initial cloud was rather insensitive to changes in the rainwater autoconversion threshold or the evaporation rate, but that the outflow temperature was somewhat warmer when the evaporation rate was decreased.

To understand the evolution of the TO, it is necessary to examine the initial cloud’s downdraft. Figure 5a shows a vertical north–south cross section of equivalent potential temperature $\theta_e$ (contour interval of 3 K) through the center of the initial cloud at 35 min (note that the vertical coordinate is stretched). The stippled region denotes the cloud downdraft ($w < -1$ m s$^{-1}$). Figure 5b shows vertical plots of vertical velocity $w$ (m s$^{-1}$), potential temperature perturbation $\theta'$ (K), water vapor mixing ratio perturbation $q_v'$ (g kg$^{-1}$), and the rainwater mixing ratio $q_r$ (g kg$^{-1}$) at $y = 29$ km (the center of the downdraft, shown by the arrow in panel a). Figure 5a shows that the cloud’s downdraft is composed of air which originates at $z \sim 2$–6 km, and that the downdraft is “fed” from the downslope direction (i.e., the cloud-relative flow is from the south in most of this region). The most intense downdraft at this time (Fig. 5b) is located at $z \sim 0.8$ km (near cloud base).

3. The control simulation

a. Characteristics of the initial cloud

In this section, we discuss results from the control simulation SA1. Since new cloud development in SA1 always occurs on the downslope edge of the advancing TO where the ambient low-level, cloud-relative flow is the strongest, the model domain is moved toward the south at a constant speed to allow for new cloud development in that region [we will use the term thunderstorm outflow (or TO) throughout the discussion of model results even though cloud electrification processes are not included in the model].

The updraft of the cloud triggered by the initial thermal impulse reaches a maximum of 34 m s$^{-1}$ at 21 min. This rather large intensity, characteristic of the first cloud (i.e., the one created by a thermal impulse)
Fig. 5. (a) A vertical north–south cross section of equivalent potential temperature $\theta_e$ (contour interval of 3 K) through the center of the initial cloud at 35 min. The stippling denotes the downdraft region ($w = -1$ m s$^{-1}$), and the D indicates air entering the downdraft at $z \approx 2$ to 3 km. (b) Profiles of vertical velocity $w$ (m s$^{-1}$), potential temperature perturbation $\theta'$ (K), the water vapor mixing ratio perturbation $q'_w$ (g kg$^{-1}$), and the rainwater mixing ratio $q_r$ (g kg$^{-1}$) through the initial cloud downdraft (position shown by the arrow in panel a) at 35 min.

and decelerates due to pressure forces as it approaches the ground. The average value of $\theta_e$ in the outflow is 327 K, which is $\approx 23$ K colder than the environment at this level.

The greatest potential temperature deficits ($-4$ to $-7$ K) occur inside the outflow near the ground (Fig. 5b). To explain the low-level temperature and moisture structure in this figure, assume that ambient environmental air at $z \approx 2$–3 km ($\theta_e \approx 327$ K) enters the storm downdraft near region D in Fig. 5a, and descends unsaturated to the ground. Air parcel trajectories (whose computation is described in section 4) indicate that downdraft air in this region does not fall vertically, but instead follows the slope of the $\theta_e$ contours shown.
in Fig. 5a. The (largely unsaturated) descent of air from region D would produce the observed \( \theta_e \sim 327 \text{ K} \) in the outflow near the surface (the only environmental air having a \( \theta_e \) of 327 K is at \( z \sim 2 \) to 3 km), with corresponding potential temperature and moisture perturbations of +8 K and -13 g kg\(^{-1}\), respectively. If \( \sim 5 \text{ g kg}^{-1} \) of rainwater are evaporated into this air as it descends near and below cloud base, \( \theta_e \) would drop from 8 K to -4 to -5 K, while \( q_c \) would increase from -13 g kg\(^{-1}\) to \( \sim -8 \text{ g kg}^{-1} \). These values agree well with the observed results inside the outflow near the ground (Fig. 5b). Although a complete thermodynamic description of the outflow air must include factors such as turbulent mixing, we believe that the transport and evaporation processes in the simplified description above are of first order importance in governing outflow thermodynamics.

### b. Collision dynamics

Although the collision of two outflows is a conceptually simple problem, the dynamics of the collision are both complex and interesting. Past studies (e.g., Mitchell and Hovermale, 1977; Goff, 1975; Purdom, 1982; and Sinclair and Purdom, 1983) have clearly shown that strong lifting is present at the gust fronts of TOs. Purdom (1982) points out that when two outflows collide, the individual vertical motion regimes of the gust fronts couple to produce a region of enhanced lifting. This region of lifting, which is typically larger than at any other location along the gust front, makes outflow collision areas favored locations for convective development. In this section, we examine the sequence of events which transpires during an outflow collision.

For purposes of illustration, consider two outflows which are several kilometers apart and are expanding along the ground in a perfectly calm environment. A nearly symmetric region of lifting (\( w > 0 \)) is present along the gust fronts as shown schematically in Fig. 6a. If we examine an area A between the two outflows, mass continuity implies that air will flow out of A as the outflows approach each other and area A becomes deformed (the outflows are \( \sim 0.9 \text{ km deep} \)). Both vertical and horizontal motions are present within A, but the largest velocities are horizontal since work must be done against gravity (i.e., static stability) to lift parcels to condensation above the neutral layer. Horizontal motion, however, requires no such work.

As the outflows move closer together (Fig. 6b), both horizontal and vertical velocities increase. (The magnitude of these velocities will of course depend on the ambient environment.) However, cloud development along the outflow CL will not begin at the point where the two outflows first meet (i.e., the center of area A). Rather, the large increase in horizontal velocity out of A creates two regions of enhanced horizontal convergence (Fig. 6b), and it is in these two regions that the first two CL clouds will form.

The model results confirm this hypothesis for a no-wind simulation discussed in Part II. However, as described in that paper, the introduction of wind shear into the base state environment augments the regions of enhanced horizontal convergence, and thus the region where the first CL clouds form. Furthermore, in settings where the outflow shapes are highly contorted, the scenario of cloud development will be more complex.

The environment along the outflow CL is strongly modified by the incipient outflow collision. As the outflows approach each other (the leading edges of the cold air being \( \sim 4-6 \text{ km apart} \)), the lifting along the gust fronts creates a region of upward motion (\( w_{\text{max}} \sim 4 \text{ m s}^{-1} \)), moistening (\( q_c \sim 2.4 \text{ g kg}^{-1} \)), and warming (due to the release of latent heat of condensation; \( \theta_{\text{e}} \sim 1 \text{ K} \)) along the outflow CL (Fig. 7). As the outflows collide at \( \sim 43 \text{ min} \), the vertical motion regimes of the gust fronts couple as discussed earlier. This is clearly illustrated for the control simulation SA1 in Fig. 8, which shows an XZ (i.e., a vertical east–west) cross section at 45 min (\( \sim 2 \text{ min after the outflow collision} \)) of \( w \) and the cold air at the point of the outflow collision (only the computational domain is pictured). Note the large difference between the updraft along the CL and the one along the eastern outflow boundary.
flows act, in a bulk sense, as quasi-impenetrable boundaries in forcing air out of their way as the collision takes place.

It is important to point out that the environment along the CL is also subject to modification due to the presence of gravity waves above the neutral boundary layer. These waves, which owe their existence to the stable region at \( \sim 870 \text{ mb} \) in Fig. 3, appear 5-10 min prior to the outflow collision and induce maximum anomalies in \( q_v \) of \( \pm 1 \text{ g kg}^{-1} \), in \( \theta' \) of \( \pm 1 \text{ K} \), and in \( w \) of \( \pm 1 \text{ m s}^{-1} \) from the undisturbed environmental values. In an experiment identical to the control simulation, but with an initial cloud separation of 26 km, we found the magnitudes of the anomalies induced by these gravity waves to be substantially smaller. However, cloud development along the CL was nearly unchanged from the control experiment. Therefore, although the anomalies induced in the control simulation may seem rather large in magnitude, they apparently have little effect on the CL convection.

To further document the dynamics involved in outflow collisions, we examine the forcing terms in the Lagrangian form of the inviscid vertical momentum equation:

\[
\frac{dw}{dt} = -C_p \vartheta \frac{\partial}{\partial z} \left( \frac{\theta'}{\theta} + 0.61q_v - q_c + q_s \right).
\]

The symbols in this equation have their conventional meaning. Equation (1) states that following a parcel, vertical velocity changes arise by virtue of vertical pressure gradient forces (VPGF) and buoyancy forces (BF). Figure 10 shows plots of the VPGF, BF, and
\( \frac{dw}{dt} \) (VPGF + BF) along a portion of the CL at 50 min (\(~7\) min after the outflow collision) for the control simulation SA1 (units are \(10^{-4}\) m s\(^{-2}\)).

The outflow along the CL is composed of rain-cooled downdraft air, and its pressure is mainly hydrostatic (i.e., vertical cross sections through the outflow show that the temperature and pressure perturbation contours are similar in shape). The highest pressure excesses are located at the ground, and as shown in Fig. 10a, the VPGF is therefore everywhere positive within the TO. Above the outflow, the values are slightly negative, reflecting the hydrostatic warming due to the release of latent heat of condensation.

The buoyancy, shown in Fig. 10b, is clearly negative everywhere inside the TO. This is simply a result of the large temperature and moisture deficits present there (\(T' \sim -5\) to \(-7\) K and \(q_e' \sim -7\) to \(-8\) g kg\(^{-1}\)). The BF is slightly positive above the TO due to condensation and associated warming there.

When the VPGF and BF are added together, we obtain \( \frac{dw}{dt} \), which is a small difference between these two large terms. The VPGF dominates the BF in the low levels near the edges of the TO, producing large regions of positive \( \frac{dw}{dt} \) (Fig. 10c). The maxima in \( \frac{dw}{dt} \) below \( z = 0.8 \) km are at the northern and southern gust fronts, as often observed in cloud simulations. As documented by Mitchell and Hovermale (1977), cold air approaching the gust front from inside the TO curves upward due to solenoidal effects as it encounters the strong baroclinic zone at the gust front. This accounts for the positive contribution to \( \frac{dw}{dt} \) there. Air outside the outflow is accelerated upward primarily by pressure forces since the horizontal temperature gradient in the ambient air is smaller than inside the outflow. Associated with both positive \( \frac{dw}{dt} \) regions is a band of horizontal convergence (typical values along the gust front in the lowest 300 m are \(10^{-2}\) s\(^{-1}\)) shown by stippling in Fig. 10c.

When a parcel enters the region of convergence above the TO, the BF takes over, with the release of latent heat of condensation providing further upward acceleration (Fig. 10c). Inside the center of the TO, \( \frac{dw}{dt} \) is weakly negative due to a nearly complete cancellation between the VPGF and BF. At the northern edge of the outflow, \( \frac{dw}{dt} \) is also positive (Fig. 10c). However, its small magnitude is insufficient in this strong wind shear profile to provide the lift needed to trigger deep clouds there.

c. Evolution of the collision line clouds

In this section we discuss the development and interaction of the clouds forming along the outflow collision line. Figure 11 depicts isochrones of the surface outflow boundary (zero degree perturbation isentrope, ground-relative framework) at 5 min intervals beginning at 30 min into the simulation. Also shown are the positions of maximum \( w \) (i.e., cell locations) relative to the ground at a height of 1.7 km above the surface. The times (in min) when the updrafts first appear at this level (which is \(~0.8\) km above cloud base) are given in parentheses beside the identifying cell letter. The lines connecting the various symbols depict the paths taken by the cells, all of which move toward the
Fig. 10. Vertical cross sections along the outflow collision line at 50 min for the control simulation of (a) the vertical pressure gradient force, (b) the buoyancy force, and (c) $dw/dt$ (all in units of $10^{-4}$ m s$^{-2}$). Solid contours denote positive values, dashed negative. The bold solid line in panel c denotes the outflow boundary ($\theta = 0$), and the stippling shows values of horizontal convergence greater than $10^{-2}$ s$^{-1}$. 
south. For clarity, the results are shown for the entire physical domain (see Fig. 2). However, in the text we will refer to results in the computational domain only.

As seen in Fig. 11 at 30 min, the outflows have just begun to spread along the ground and are nearly circular in shape. However, the shear above cloud base causes the rainy downdraft of the initial cloud to tilt toward the south (see section 3a), producing a rain area at the surface which is elongated in the north–south direction. Since the motive force for TOs is primarily the hydrostatic pressure excess in the rain-cooled air (e.g., Wakimoto, 1982), the TOs reflect the asymmetry of the rain region and also become elongated in the north–south direction (similar to Thorpe and Miller, 1978). The overall outflow speed remains fairly constant despite the large number of convective cells forming and decaying along the gust front.

The first collision line cell CL1 (the convention for naming these cells is that CL1 refers to the first cloud to form along the outflow collision line, CL2 the second, etc.) appears at 1.7 km at 50 min (Fig. 11), and attains a maximum $w$ of 23 m s$^{-1}$ at 65 min. The updraft of CL1 rises vertically like a buoyant thermal, and is located above the 1.7 km level by 65 min. A
second cell CL2 is triggered just after and to the south of CL1 (note from Fig. 11 that CL2 is behind the surface outflow position by 55 min). Cell CL2 is \( \sim 6 \text{ m s}^{-1} \) weaker than CL1, primarily because CL2 forms beneath the strong, downshear-tilted downdraft of CL1 (see section 4). Although the rain-cooled downdrafts from these CL cells penetrate and reinforce the temperature deficit in the outflow, there are no "bulges" or localized regions of enhanced horizontal convergence anywhere along the gust front. Recalling Fig. 6, which implied that two updrafts will be triggered when two outflows collide, we can see that CL1 and CL2 reach their maximum intensities within 5 min of each other because they are triggered mainly by the aforementioned forced uplifting along the outflow collision line.

A third cell CL3 \( (w_{\text{max}} = 28 \text{ m s}^{-1} \text{ at 84 min}) \) is triggered along the advancing TO, and reaches the 1.7 km level at 70 min. It is interesting that CL3 is not only the most intense of all CL cells in this simulation, but also that the time between the maximum \( w \) of CL3 and CL2 is about three times longer than that between CL1 and CL2. As discussed in the section 4, this is a result of the fact that CL1 and CL2 are triggered primarily by the forced lifting of air due to the outflow collision, while CL3 forms as air is forced to rise up and over the advancing gust front. In addition, unlike CL2, the development of CL3 is not impeded by downdrafts to its north (section 4).

Although the focus of this paper is the convection triggered along outflow collision lines, it is instructive to also note the characteristics of the clouds triggered off the CL (i.e., south of the initial cloud). As will be shown in Part II, several similarities exist between convection on and off the CL in various wind shear regimes.

Figure 11 shows that cells B and B', located south and to the side of the initial cloud A, are longer-lived than any of the other CL cells. B and B' reach maximum intensities of \( \sim 24 \text{ m s}^{-1} \) at 72 min, and vertical cross sections through their updrafts indicate that they are strongly rooted in the boundary layer. It is important to point out that cells B and B' are discrete entities triggered along the sides of the expanding outflow, and that these cells are not a result of a split by cell A (though a split did occur in a similar situation reported by Thorpe and Miller, 1978). At 70 min, cells C and C' reach the 1.7 km level south of B and B', with cell C' being the most intense. Thereafter, only relatively weak \( (w_{\text{max}} \sim 10 \text{ m s}^{-1}) \), shallow clouds form in the unstable airmass along the advancing TO.

It is interesting that no clouds are ever triggered along the gust front directly south of the initial cloud (Fig. 11). The explanation for this lies in the strong environmental wind shear, which causes the updraft, downdraft, and rain region of the initial cloud to tilt significantly in the downshear direction (Fig. 12). As the rain falls, it warms by compression \((\theta' \sim 0.5-1.0\text{ K})\) the rain-free region beneath it, which is located ahead of the gust front. Accompanying this warming is subsidence \((w < -0.5 \text{ m s}^{-1})\) and associated warming \((\theta' > 0.5 \text{ K})\) due to the downshear tilt of the downdraft. The outflow \((\theta' < 0 \text{ K})\) is stippled.

![Fig. 12. A vertical north-south cross section through the initial cloud in the control simulation SA1 at 65 min showing the subsidence \((w < -0.5 \text{ m s}^{-1})\) and associated warming \((\theta' > 0.5 \text{ K})\) due to the downshear tilt of the downdraft. The outflow \((\theta' < 0 \text{ K})\) is stippled.](image-url)
finer grid resolution and improved numerical techniques are underway to explore the nature of these waves, particularly with regard to their role in convective cloud development.

The collision of the outflow with its mirror image counterpart is clearly evident at 50 min. Note how the depth of the cold air increases along the CL as a result of the collision. This elevated region decreases somewhat in height as the outflow spreads along the ground until, by 90 min, the TO has become more horizontally uniform due to the absence of any significant downdrafts penetrating its upper surface.

4. Analysis of the control simulation from air parcel trajectories

Air parcel trajectories are a useful tool for examining the flow structure of TOs. Mitchell and Hovermale (1977) and Thorpe, Miller, and Moncrieff (1982) used trajectories in their two-dimensional numerical models to trace air parcels in and around simulated TOs. Their results have advanced the knowledge of outflows, particularly with regard to providing dynamical interpretations of their associated flow structure.

Based on data obtained from instrumented aircraft penetrations, Sinclair and Purdom (1983) constructed a vertical cross section through an advancing TO (Fig. 15). Note the general subsidence (A) far ahead of the gust front, and the broad region of strong upward motion (B) closer to the TO. Also indicated in this figure is the mixing present in the gust frontal zone (e.g., Simpson, 1972). Finally, note that in this diagram air approaching the TO over a deep layer is shown to rise into the cloud arc line. One must, however, consider
the total three-dimensional picture since air parcels can pass around, as well as up and over, an advancing TO. In addition, one would not expect all of the air rising up and over the outflow to continue upward into the cloud arc line.

To help clarify the flow structure associated with the modeled TOs, we now present results of fully time-dependent (nonsteady state) kinematic air parcel trajectories which are allowed to move through the modeled clouds. Three points are addressed in this section: The first deals with the motions of air parcels along the CL near the time of the outflow collision, and the subsequent development of the first collision line cloud CL1. Second, we focus on the trajectories of air parcels approaching the gust front to elucidate the multiple cell development along the CL. Finally, we use trajectories to determine the origin of air in the updrafts of cells CL1 and CL3 (recall that these two cells are triggered by different mechanisms).

In computing an air parcel trajectory interactively from model data fields stored at 5 min intervals, we specify the time and starting location of the parcel. The local three-dimensional wind field is interpolated to that point, and then the parcel is extrapolated forward and backward in space (no more than 100 m per integration) and time. The wind field is then interpolated to the parcel’s new location, and the computations are repeated. Since each parcel extrapolation is computed using the most recent wind data, these are truly time-dependent air parcel trajectories.

Due to the transient nature of the modeled clouds in these experiments, we conducted tests to determine the sensitivity of the trajectory path to the parcel’s initial location. In general, a small change (≈0.5 km) in the parcel’s initial location produces only a minor quantitative change in the parcel’s trajectory. We also computed selected trajectories from model data fields stored at 1 min intervals, and compared them with corresponding trajectories computed using the 5 min data. We found that, although some individual trajectories were different between the two data sets (particularly in the rapidly changing cloud region), the basic flow structure was the same. Therefore, trajectories from the 5 min data set are utilized here. Of the several hundred trajectories computed during the course of these experiments, only a few are presented to illustrate

FIG. 15. A vertical cross section through a thunderstorm outflow boundary which is moving from left to right. The data were obtained from several penetrations by an instrumented aircraft. “A” denotes the area of subsidence well ahead of the gust front, while “B” marks the position of strong upward motion at the gust front. (From Sinclair and Purdom, 1983).
the points of this section. However, they are felt to be representative of the flow of many similar parcels in the regions described.

The starting location for each trajectory shown is within the model symmetry plane (i.e., along the outflow CL), and thus each parcel will remain in that plane since $u = 0$ there. This implies that all parcels must pass up and over the TO. Although this may seem to be an unnecessary and unrealistic restriction in a three-dimensional model, CL clouds in the high shear control simulation to be examined here receive almost all of their air from the downshear direction (i.e., air that rides up and over the TO), and only a small portion from lateral flow. Trajectories of air parcels placed 1–3 km east or west of the CL exhibit similar behavior to those which are constrained to remain in the CL; i.e., they all flow into the CL clouds. Due to the strong ambient winds and irrotational nature of these clouds, trajectories placed further than 3 or 4 km on either side of the CL simply pass by the CL clouds.

a. Trajectories near the time of the outflow collision

To obtain a general idea of the airflow along the CL before, during, and after the time of the outflow collision, we place air parcels in the flow field at various heights and times in the control simulation SA1. The starting locations of the parcels are shown in Fig. 16. At 35, 40, and 45 min, 7 parcels are initiated in a vertical stack at $y = 31$, 34, and 37 km, respectively, at the heights shown in the figure. The center of the outflow collision (the point at which the two outflows first meet) is also indicated.

Figures 17a–c show the trajectories of parcels initiated at 35 min (~8 min prior to the outflow collision). The wide arrows enclose the paths taken by these parcels, while the lines through the arrows indicate the time in min along the trajectory paths. Trajectories in this figure are displayed using the vertically stretched grid employed in the model. All parcels approaching the collision region at this time (Fig. 17a) move upward due to the lifting in that area by the outflow collision. By 65 min, these parcels are rapidly accelerated upward inside of cell CL1. Parcels begun at $y = 34$ km (Fig. 17b) have passed to the north of the collision region by 45 min. All parcels eventually rise but then descend to the north of the developing cell CL1. Note how parcels on the bottom side of the arrow are rapidly accelerated northward and upward from 45 to 55 min due to forcing by the outflow collision. Parcels initiated at $y = 37$ km (Fig. 17c) are seen to simply move northward with the ambient flow from 35 to 45 min, with rather large accelerations in the low levels after the collision at~43 min.

Moving to 40 min, or ~3 min prior to the outflow collision, we see a pattern similar to that at 35 min for parcels starting at $y = 31$ km (Fig. 17d). Again, all parcels enter updraft CL1 and are rapidly accelerated after 50 min. Note that the parcels slow down from 60 to 70 min as they make the sharp southward turn in the upper levels of CL1 under the influence of the strong upper level winds. At $y = 34$ km (Fig. 17e), only the top few parcels in the stack reach the upper levels of the model atmosphere. The rest begin to rise, but at 60 min are north of the main updraft and are forced downward by compensating subsidence associated with CL1. Parcels released at $y = 37$ km (Fig. 17f) exhibit the same behavior as those at 35 min. Again, note the rapid acceleration of the lower parcels in the stack after 40 min due to forcing by the outflow collision.

At 45 min, or ~2 min after the outflow collision, the trajectory patterns are somewhat different. At $y = 31$ km (Fig. 17g), only parcels near the ground enter updraft CL1. Note the rapid acceleration of these parcels from 55 to 65 min. Parcels higher in the stack rise more quickly at first (and to the south of those that rise higher), but they soon encounter compensating subsidence associated with CL1, and by 70 min, the strong rainy downdraft of CL1 forces them downward. Parcels at $y = 34$ km (Fig. 17h) are accelerated rapidly upward into CL1 after the outflow collision. However, the bottom few parcels, which were in the cold outflow when the integration began, rise only a few hundred meters under the influence of the outflow collision forcing, and then sink due to insufficient buoyancy relative to the ambient air. At $y = 37$ km, the trajectories exhibit the same behavior as those at 35 and 40 min.

b. CL trajectories approaching the outflow

To examine the nature of the ambient airflow approaching an advancing TO, we present trajectories of air parcels which are placed at various heights slightly ahead of the outflow boundary. In order to obtain the most complete picture of the flow field, each trajectory is integrated both forward and backward in time from its specified starting location.

Fig. 16. A schematic vertical cross section along the CL showing the starting positions of the air parcels in the trajectory computations. Parcels are released in these three stacks at 35, 40, and 45 min.
Figure 18 shows vertical cross sections along the outflow CL of $w$ (thin solid lines, contoured at 1, 4, 10, and 20 m s$^{-1}$), the cold air outflow boundary ($q' < 0$ K, bold lines), and air parcel trajectories (solid lines). These cross sections are plotted in the vertically stretched coordinate system, and thus the subcloud layer appears much deeper than it actually is in comparison to mid- and upper-levels. Only the bottom and right sides of each figure panel correspond to computational boundaries in the cloud model. Panels are shown for every 10 min, and the parcel location indicated by a letter corresponds to the time shown in the panel. The bold solid lines show the movement of each parcel over the period 5 min prior to the time shown in the figure panel. The model domain is moved at approximately the speed of the gust front ($\sim 12$ m s$^{-1}$), and thus the parcel trajectories are nearly outflow (storm) relative.

The first three parcels (A, B, and C) are placed at $z = 0.1$, 0.3, and 0.7 km, respectively, at 45 min (approximately 2 min after the outflow collision). Parcel A is within the cold air near the surface, while B is at
the cold air interface. Parcel C is located well above the cold air at this time.

At 50 min (Fig. 18a), parcel A is rising inside the TO (due to lingering upward motion in the cold air from the outflow collision), while B and C are both well outside of the cold air pool. It appears that parcel B has crossed the upper interface of the TO, when in fact it rises almost coincidently with the outflow interface. This visual discrepancy arises from viewing a continuous process at finite time intervals. Between 45 and 50 min, the outflow depth increases from 0.5 to 0.9 km (as Fig. 14 shows qualitatively). Note from Fig. 18a that the region of lifting ($w > 1 \text{ m s}^{-1}$) caused by the outflow collision is largest in size toward the southern edge of the cold air pool.

By one hour (Fig. 18b), two distinct updrafts are present along the CL. Note that parcel A has apparently passed through the cold air boundary, and is in the strongest portion of CL1's updraft. Parcels B and C, which came from higher in the boundary layer, are near the edge of updraft CL1. It is interesting to note that although CL1 has moved several kilometers behind the gust front, it continues to grow. Parcel trajectories initiated near CL1's updraft show that it is sustained primarily by air which enters the updraft near cloud base after riding up and over the TO (e.g., parcel C), and to only a small extent by lateral airflow.

At 55 min, four parcels are inserted and followed forward and backward in time ($A', B', C'$, and $D'$ at $z = 0.1, 0.3, 0.7, \text{ and } 1.1 \text{ km}$, respectively). These parcels move nearly horizontally as they approach the gust front. However, in the region 1–2 km ahead of the outflow (Fig. 18b), they turn upward rather abruptly. The lowest three parcels ($A', B', \text{ and } C'$) actually cross the cold air interface and travel as much as 3 km into the cold air pool (a similar behavior was observed using
trajectories computed from model data stored at 1 min intervals). Turbulent mixing at the gust front, schematically illustrated in Fig. 15 and present within the model, appears to be the major factor in the development of this cross-zone flow.

At one hour (Fig. 18b), parcels B', C', and D' have been captured by CL2, and are thereby effectively blocked from reaching CL1. Of course, CL1 could obtain air laterally from off the CL, but as stated earlier, such flow is insignificant in this wind shear profile. Therefore, in addition to precipitation loading in its updraft, CL1 decays as the outflow outruns it because low-level, unstable air is effectively “cut off” by new cells forming at the gust front (similar to Williamson and Chen, 1982).

The new cell CL2 is fed by parcels B', C', and D' as illustrated in Fig. 18c. However, unlike parcel A which crossed out of the TO and entered cell CL2, parcel A' simply rises inside the upper portion of the TO where mixing prevents it from entering CL2s updraft. Cell CL3 is beginning to grow at the gust front, and has an updraft of 4 m s⁻¹ at this time. Four more air parcels are initiated at this time (A'', B'', C'', and D'') at the same height as the previous four. Tracking these parcels backward in time, we find that as they approach the gust front, they exhibit a general tendency to rise by 50–100 m in the region 4–8 km ahead of the TO.

One final observation from Fig. 18c is that of waves that develop in the upper portion of the cold outflow in response to the precipitation and downdraft air which penetrate the TO (see Fig. 14 and section 3c). These waves are not a result of numerical instability. Rather, they appear to propagate along the top of the TO as gravity waves. Their importance in the dynamics of this type of convection requires further investigation.

By 80 min, CL2 is decaying as the downdrafts from both CL1 and CL2 penetrate the TO, initiating more waves at the upper edge of the cold air pool (Fig. 18d). The large depression in the TO interface is just north of the surface rain region. Cell CL2 never reaches the strength of CL1 because of its proximity to both low and upper level downdrafts associated with CL1. Cell CL3, however, forms approximately 15 min after CL2 and is more intense. Furthermore, in contrast to cells CL1 and CL2, which are both rather quickly outrun by the gust front, cell CL3 remains at the gust front for nearly its entire lifetime.

Parcels B'', C'', and D'' make their way inside of CL3s updraft, with B'' passing through the upper leading edge of the cold outflow (Figs. 18c and d). It is clear from this figure that cell CL3 is still growing at the gust front (i.e., in the region of maximum convergent forcing) and is relatively uninhibited in its growth by downdrafts to the north.

By 90 min (Fig. 18e), parcels B'', C'', and D'' are located in the upper portion of CL3. Meanwhile, parcel A'', which originated near the surface from within the outflow, is subsiding inside of CL3s downdraft.

It is interesting that some parcel trajectories pass through more than one convective cell. For example, B'' and C'', which were within updraft CL2 at 65 min (not shown), have now become a part of updraft CL3. From these trajectories and others not presented here, it is apparent that air parcels from near the surface can enter one updraft, and then exit to become part of another updraft. This recirculation of air is similar to that of hail embryos and graupel in severe multicellular storms (e.g., Heymsfield et al., 1980).

c. CL updraft trajectories

This section focuses on the trajectories of air parcels released within the updrafts of the CL clouds. By following these parcels forward and backward in time, we may identify the origin of air supplying the simulated clouds.

Recall from section 3 that CL1 and CL2 are triggered nearly simultaneously by the forcing of the outflow collision. In contrast, CL3 is initiated much later at the leading edge of the TO by air that rises up and over the gust front. Given these two different forcing mechanisms, the natural question to ask is whether the properties of the updraft air in CL3 are significantly different from those in CL1 or CL2. To address this question, we compute trajectories of air parcels placed at every grid point within updrafts CL1 and CL3. In order to make a meaningful comparison, the cell updrafts are required to be in similar stages of growth. To satisfy this requirement, we begin the parcel integrations at 60 min for CL1 and at 80 min for CL3.

Figures 19a, b show vertical cross sections along the CL through updrafts CL1 at 60 min and CL3 at 80 min, respectively (contour interval is 2 m s⁻¹). The stippled areas denote those regions where air originates at or below 0.3 km. From this figure we see that CL1 and CL3 are composed of air primarily from very near the ground. Furthermore, the updraft cores of these cells are virtually undiluted. We have confirmed this result thermodynamically by examining values of θ within the modeled clouds. This result agrees with an observational analysis by Davies-Jones (1974), who found that strong updraft cores which reach midlevels are typically undiluted.

Since cells CL1 and CL3 are composed of the same type of air, and since the CL cells are fed mainly by low-level air from the south that is forced to rise up and over the TO, the timing difference between CL1 and CL2 (5 min) versus that between CL2 and CL3 (~15 min) is a result of the outflow collision. As the two outflows collide, they produce a large region of lifting along the CL (see Fig. 7). Two updrafts (CL1 and CL2) are triggered by this lifting process with the upshear one (CL1) being the strongest in the control case (see Fig. 18b). Note from this figure that since CL1 is triggered by the outflow collision, it is already well behind the leading edge of the TO early in its
growth of CL2 as shown schematically in Fig. 20b and discussed for the initial cloud in section 3a. This suppression effect could occur for any pair of cells growing a few kilometers apart in a highly sheared environment (Turpeinen, 1982).

In contrast to CL1 and CL2, cell CL3 is triggered away from the outflow collision at the leading edge of the TO by air rising over the gust front. Thus, CL3 is neither influenced by the outflow collision nor by potentially detrimental tilted downdrafts to the north. CL3 essentially behaves like a cell triggered at the gust front of a single storm. Cells which subsequently form along the CL behave like cells forming along a single TO.

5. Impact of low-level moisture changes on the control simulation

a. Discussion of cell development

Up to this point, the ambient environmental temperature, moisture, and wind profiles have been held fixed in the control simulation. We now examine the character of cloud development for the control simulation wind shear profile when the low-level (surface to ~1 km) moisture (LLM) is increased or decreased. All parameters except for the moisture profile remain the same.

Figure 21 shows the lower portion of a SkewT–logP diagram for the three varied moisture simulations, along with the control case SA1. In the two "dry" experiments SA1DRY1 and SA1DRY2, the LLM is decreased from ~1000 to 900 mb. In the "moist" case SA1MOIST, the LLM is increased in the shallower layer from ~1000 to 950 mb. Since the characteristics of simulated clouds are known to strongly depend on the LLM (e.g., Weisman and Klemp, 1982), we felt it appropriate to confine our alterations in the moisture to the boundary layer. This is not meant to be an exhaustive test of the behavior of CL clouds to changes in the LLM. Rather, we merely wish to identify the tendencies of these simulated clouds when such changes take place.
Table 2 lists the maximum and minimum updraft statistics (m s\(^{-1}\)) for the control experiment and for the varied moisture simulations. The times in minutes when the extremes in \(w\) are reached are shown in parentheses. The strength of the cloud triggered by the initial thermal impulse is clearly a function of the LLM. When the LLM is decreased, this cloud weakens substantially, and more time is needed to reach its peak intensity. The variation in downdraft strength among the experiments is much smaller than that of the updrafts, and the time delay due to the moisture variation is also evident.

Turning to the CL clouds, Table 2 shows that several strong convective cells are triggered along the CL in rapid succession in the moist simulation. The intensities of the first three cells are quite similar to those in SA1. However, note that CL1 in the moist run reaches its maximum \(w \sim 6\) min earlier than in the control case. For the second cells, the difference is only 3 min, while for CL3 there is an 8 min difference between the time \(w_{\text{max}}\) is reached in SA1MOIST compared to the control case. In addition, SA1MOIST is the only simulation where a strong fourth CL cloud is present by the end of the simulation.

Looking at the decreased moisture experiments in Table 2, we see that a decrease in moisture not only weakens all CL clouds, but also increases the time between successive cell development. In the driest case SA1DRY2, cloud development is almost completely suppressed along the CL.

In contrast to the control simulation, vertical cross sections through the CL clouds in SA1MOIST (e.g., CL2) show that new updrafts are able to grow into tilted downdrafts without suffering any appreciable suppression effects (compare Fig. 20b). This is a result of the increased LLM and associated updraft enhancement, as well as the fact that the cloud-to-cloud spacing along the CL in SA1MOIST is somewhat larger than in the control case.

To obtain a better idea of the impact that an increase in LLM induces on cell formation relative to the control case, a horizontal cross section at \(z = 1.7\) km for simulation SA1MOIST is shown in Fig. 22. Comparing this with the control run in Fig. 11, we see that an increase in LLM encourages the rapid development of many clouds all along the TO. However, note that as in the control simulation, no significantly strong cells form directly south of the initial cloud A. The bulge along the gust front south of cell A in Fig. 22 is more pronounced than in the control case (Fig. 11). This bulge is a reflection of the tilted rainy downdraft of cell A, and since the magnitude of this cell’s downdraft in SA1MOIST is larger in strength and aerial extent than in the control case, the bulge is correspondingly more evident in SA1MOIST. Note, however, that by one hour in both simulations, this bulge has effectively disappeared.

It is interesting that cells on the east side of the expanding TO are conspicuously absent even when the LLM is increased. This implies that no deep clouds would have formed to the west in the absence of the outflow collision.

### b. Effects of low-level moisture on the outflow

Changes in the LLM affect the TO as well as the developing clouds. The maximum outflow temperature deficits in the moist run (and in the control run) are \(\sim 3-4\) K colder than in the two dry simulations (the outflow depths are virtually the same in all experiments). This causes an \(\sim 3\) m s\(^{-1}\) increase in the outflow speed over the dry runs.

### Table 2. Maximum updraft and downdraft statistics (m s\(^{-1}\)) for the varied moisture experiments.

<table>
<thead>
<tr>
<th>Experiment</th>
<th>Initial (w_{\text{max}})</th>
<th>Initial (w_{\text{min}})</th>
<th>(w_{\text{max}}) CL1</th>
<th>(w_{\text{max}}) CL2</th>
<th>(w_{\text{max}}) CL3</th>
<th>(w_{\text{max}}) CL4</th>
</tr>
</thead>
<tbody>
<tr>
<td>Low-level</td>
<td>SA1MOIST</td>
<td>37 (21)</td>
<td>-14 (29)</td>
<td>25 (59)</td>
<td>21 (66)</td>
<td>28 (76)</td>
</tr>
<tr>
<td>Moisture</td>
<td>SA1 (control)</td>
<td>34 (21)</td>
<td>-13 (30)</td>
<td>23 (65)</td>
<td>17 (69)</td>
<td>28 (84)</td>
</tr>
<tr>
<td>Increasing</td>
<td>SA1DRY1</td>
<td>30 (22)</td>
<td>-11 (32)</td>
<td>16 (66)</td>
<td>10 (&gt;90)</td>
<td></td>
</tr>
<tr>
<td></td>
<td>SA1DRY2</td>
<td>25 (26)</td>
<td>-10 (32)</td>
<td>6 (70)</td>
<td></td>
<td></td>
</tr>
</tbody>
</table>
\[ dV_0 = C d\theta / (\phi')^{1/2}, \]  

where \( C = 0.5k (gH/\bar{\theta})^{1/2} \). For the outflows in this study, \( C = 2.4 \text{ m s}^{-1} \text{ K}^{-1/2} \). Using the values for \( \theta' \) and \( d\theta' \) from the model results for SA1MOIST and SA1DRY2 in (3), we find that \( dV_0 = 2.9 \text{ m s}^{-1} \). This result agrees well with the observed 3 m s\(^{-1}\) difference in outflow speed even though (3) does not include surface frictional effects.

The explanation for the colder outflow temperatures in the “moist” and control experiments lies in the fact that, as a whole, cells in these simulations have stronger downdrafts (~20–40% stronger) and produce larger quantities of rain (an order of magnitude more) than in the “dry” simulations. The evaporation of this rain by the downdrafts produces colder temperatures inside the outflow. In contrast, the dry simulations have less moisture available for rain production, therefore producing less evaporation for cooling of the outflow. This causes the outflow temperatures to be somewhat warmer. It is interesting to note that an increase in LLM has little effect on the outflow temperature, while a decrease in the LLM induces significantly warmer temperatures inside the outflow.

6. Discussion and summary

The development of convection along intersecting thunderstorm outflow boundaries was investigated using a three-dimensional numerical cloud model. The configuration of the experiments was deliberately chosen to be simple in order to extract as much physical information as possible from a complex problem.

The model initial environment for the control simulation was characterized by temperature and moisture profiles typical of strong convective situations. The model initial wind field was prescribed by a constant unidirectional shear of 2.9 m s\(^{-1}\) km\(^{-1}\) from 0.8–8.9 km, with a constant wind everywhere else. In each simulation, two mirror image initial clouds were triggered by artificial thermal impulses located at the top of the boundary layer. The line containing the two initial clouds (which were spaced 16 km apart) was oriented perpendicular to the wind shear vector.

As the precipitation-induced outflows from the two initial clouds approached each other, they induced lifting, warming, and moistening in the region where they eventually intersected (the outflow collision line, or CL). Time-dependent air parcel trajectories and thermodynamic data indicated that the outflows literally “squeezed” air out of their way as they approached each other and collided. Two clouds were triggered along the CL due to the outflow collision. The upshear member of this pair of clouds had a slight head start in development, and since the clouds were growing close together and competing for the same air, the upshear cloud became the strongest (the downshear cell reached its maximum updraft intensity ~5 min after the first). In addition, the downshear cell was sup-
pressed because it grew into the region occupied by the upshear cell's downdraft.

As low-level air parcels along the CL approached the outflow from the south, the vertical pressure gradient force of the outflow deflected them upward over the gust front. This frontal lifting triggered a third cell along the outflow CL. In contrast to the first two clouds, which were triggered by the forcing of the outflow collision, the third cloud was triggered by air that was forced to rise up and over the advancing gust front. This was supported by the fact that the third cell reached its maximum updraft intensity ~15 min after the second cell, and was the most intense of the three clouds which formed along the outflow CL. This larger time interval also meant that the gust front had moved away from the region of the first two clouds when the third was triggered. Thus, the growth of the third cloud was not hampered by tilted downdrafts from clouds upshear. Air parcel trajectories also indicated that even though the first two CL clouds were triggered by different mechanisms than the third, the air within the updraft "cores" of these clouds was virtually undiluted, originating at or below 0.3 km.

Clouds along the CL were fed mainly by low-level air which was forced to rise upon encountering the gust front. The lateral flow into the CL clouds was relatively small. The decay of cells along the CL was governed partly by precipitation loading in the cloud updrafts, and partly by new cells forming at the gust front which effectively blocked low-level air from reaching and sustaining previous cells.

Some low-level air parcels approaching the outflow from the south crossed the periphery of the gust frontal boundary. This was believed to be a result of turbulent mixing at the gust frontal zone. Once above the outflow, the air parcels were found to pass through several updrafts and downdrafts as they traversed the cloud region.

We stress that the trajectories presented here are purely kinematic; no thermodynamic information was used in their computation. Therefore, there exists some uncertainty as to the accuracy of these trajectories near large thermal gradients, e.g., parcels that cross the cold air interface. If parcels do indeed cross the interface as observed in this and other simulations not discussed here, this raises the fundamental question of how one should really define an outflow boundary (e.g., by temperature, density, winds). It is likely that numerical models used to study outflows, including the one used here, have insufficient resolution and too much inherent numerical diffusion to accurately give more than a qualitative picture of the outflow interface (whatever one defines it to be) and trajectories near it.

The modeled clouds were found to be sensitive to changes in the low-level moisture (LLM). Increasing the LLM caused all clouds in the simulation to become stronger, and also caused them to develop more rapidly than in the control simulation. Decreasing the LLM had the opposite effect, to the point that only shallow clouds formed along the CL. Furthermore, a decrease in the low-level moisture decreased the temperature deficit inside the outflow, thereby reducing its speed. This result was shown to be consistent with classical inviscid density current theory.

It could be argued that results similar to those obtained by varying the LLM could be achieved by changing the microphysical parameterization (e.g., statistical distribution of raindrop sizes, evaporation rate, autoconversion threshold, etc.). However, we found that for our model, the results tended to be controlled more by changes in the ambient environment than by changes to the microphysics.

We wish to point out that this study is part of a long range effort aimed at obtaining a better understanding of the many roles of thunderstorm outflow boundaries in the life cycle of convection. Some of the major questions concerning outflows which remain unanswered are

1) How does surface friction affect an outflow and the convection triggered by it?
2) What factors determine whether air approaching an outflow boundary will rise up and over, or simply pass around the cold air pool?
3) How do low-level static stability and vertical wind shear affect the motions of air parcels approaching the gust front?
4) What determines the depth of an outflow?
5) Under what conditions would one expect to forecast strong or deep outflows?
6) What factors determine where along an advancing outflow boundary deep convection will form?
7) What is the relative importance of rain and cloud water evaporation, mixing, and advection in producing the temperature and moisture profiles in observed and simulated outflows?

Several of these questions are being addressed with the three-dimensional model and with a newly-developed two-dimensional outflow model.

In order to substantiate our model results, more observations of outflows are needed. Satellites are a useful tool for tracking outflow boundaries that are not obscured by clouds. Studies relating satellite-observed outflow movement to storm and environmental characteristics would be of great importance. Ground-based and airborne Doppler radar observations would allow reconstruction of the three-dimensional wind field of the outflow. In addition, such observations could better identify the regions of turbulence and mixing associated with the outflow, and comparison could be made with laboratory and numerical model results. Finally, multiple aircraft penetrations through outflows and observations by ground-based intercept teams (e.g., time lapse movies and remote rawinsonde releases) would
provide further valuable data on the structure of thunderstorm outflows.

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