An Explanation for Intense Frontal Updrafts and Narrow Cold-Frontal Rainbands

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ABSTRACT

Measurements with Doppler radar, and instrumented aircraft and towers, have revealed that surface cold fronts often have cross-frontal circulations organized on a scale of a kilometer or less. These circulations include intense updrafts (1 to 20 m s^{-1}) that result in a narrow band of heavy rainfall. We used a nonhydrostatic model to investigate the mechanism for these updrafts and to isolate those characteristics of the prefrontal environment that result in intense updrafts and narrow bands of heavy rainfall. Our simulations were initialized with a cold reservoir in a manner analogous to that used to produce a gravity current. The similarity between the observations and our simulated frontal flows supports the hypothesis that the flow at the leading edge of surface cold fronts can sometimes be represented by gravity-current dynamics. We also found that the differences between frontal circulations and classical dry gravity currents can be explained by the effects of precipitation and vertical shear. In our simulations, the intense updrafts at the leading edge of the cold air mass were associated with a strong upward-directed pressure force and were not associated with significant parcel buoyancy. The conditions for intense updrafts and heavy rainfall in our simulations were 1) strong deep cold pools, 2) a prefrontal environment that contains deep layers of air that are nearly saturated with a lapse rate that is nearly neutral to moist ascent, and 3) intense low-level vertical shear in the cross-frontal component of the horizontal wind. These conditions are typical of maritime surface cold fronts that often have strong updrafts and narrow bands of heavy rainfall. In our simulations, the vertical shear in the cross-frontal direction exerted a strong influence on the strength and character of the frontal updraft. For a given magnitude and depth of the cold air an optimal vertical shear existed where the updraft was vertical, intense, and resulted in a narrow band of heavy precipitation. With decreasing vertical shear, the updraft tended to weaken, tilt back over the cold air, and result in a broad band of lighter precipitation. An unsteady system resulted at shears higher than optimal. The dependence of the updraft character on vertical shear is similar to that predicted by recent theoretical work on squall lines.

1. Introduction

During the past three decades, advanced technology such as Doppler radar and instrumented aircraft and towers afforded a new look at cold-frontal circulations. The early studies of surface cold fronts with Doppler radar (e.g., Browning and Harrold 1970, 1971; Browning 1971, 1974; Browning and Pardoe 1973) found that 1) the cross-frontal gradients in the lowest 1–2 km of height were often concentrated into regions of a kilometer or less, 2) shallow updrafts at the leading edge of these fronts had magnitudes in excess of a few meters per second, and 3) a line of heavy precipitation was located along the front (often termed a narrow cold-frontal rainband or line convection). These strong updrafts were found to occur despite a lack of appreciable potential instability in the warm sector. Subsequent high-resolution measurements of other surface cold fronts (e.g., Carbone 1982; Hobbs and Persson 1982; Parsons and Hobbs 1983a,b) clearly show that narrow widths and strong, shallow updrafts are commonly associated with the leading edge of well-defined cold fronts in middle latitudes. Of these studies, the Carbone (1982) system, with its updrafts locally ranging up to 20 m s^{-1}, heavy rainfall, and a weak tornado, is particularly noteworthy, since it also illustrates that these fronts can be associated with severe weather even in the absence of appreciable potential instability.

The flow associated with the surface fronts observed in these studies appeared similar to convective outflows. However, their origin appeared not to be of a convective nature, since researchers found that the fronts clearly marked the boundary between two distinctly different synoptic-scale air masses and the systems were observed in maritime environments that were nearly saturated and convectively neutral and, therefore, not conducive to evaporatively driven convective downdrafts and outflows. Further evidence to reject the convective outflow hypothesis was found by Shapiro (1984) and Shapiro et al. (1985), who showed that the narrow horizontal scale in the cross-frontal direction and accompanying updrafts also occurred in fronts lacking significant clouds and precipitation. It should
be noted that these finescale fronts can also occur apart from any orographic influences, as demonstrated by the Bond and Fleagle (1985) study of a front over the Pacific Ocean. Although one might argue the bias of scientists toward investigating only the more severe events, these high-resolution studies of fronts have typically not revealed surface fronts to be broad, uniform, transition zones, further suggesting that these finescale circulations may be intrinsic to well-defined cold fronts.

One should note that, although these observations represent a return to the view that surface fronts represent near discontinuities in wind and temperature, these recent observations do not suggest a return to the Margules (1906) view of fronts where the flow associated with these discontinuities is in balance with rotation. Instead, numerous studies (e.g., Carbone 1982; Hobbs and Persson 1982; Shapiro 1984; Shapiro et al. 1985) have suggested similarities between these sharp fronts and laboratory and atmospheric gravity currents (Simpson 1969, 1972). Thus, the theoretical insight distilled from these high-resolution studies is that the gradients in the cross-frontal direction can sometimes become large enough so that rotation is of secondary importance, and the front behaves as a gravity current with a strong updraft that is significantly nonhydrostatic and winds locally ageostrophic in the across- and even alongfrontal directions (e.g., Parsons et al. 1987). While one might be tempted to conclude from these studies that strong fronts that behave as density currents will contain intense updrafts, theoretical studies into squall-line dynamics (e.g., Thorpe et al. 1982; Rotunno et al. 1988) showed that the component of the ambient vertical shear of the horizontal wind normal to the front provides a strong constraint on the behavior of updrafts at the leading edge of a gravity current.

This study assumes the existence of these sharp cold fronts and numerically investigates which factors in the warm-sector environment influence the updraft at the leading edge of the front. Our work explores the application of our knowledge of convective and squall-line theory (for a review see Rotunno et al. 1988) to the near neutrally buoyant updrafts observed in the previously mentioned frontal cases. Our findings relate the Rotunno et al. interpretation of the strong dependence of the intensity of frontal updrafts upon the balance between the vertical shear and the “strength” of the cold air mass to these frontal systems. We believe that the simulations presented here are the first high-resolution simulations of these intense frontal updrafts. The agreement between our simulations, initialized with a cold reservoir in a manner analogous to that used for gravity currents, and observed fronts suggests that the circulations associated with the leading edge of surface fronts are well represented by gravity-current dynamics. While a previous study by Smith and Reeder (1988) had used differences between observed flows and steady “classical dry gravity currents” to discount the gravity-current analogy, we propose that many of these differences are simply the result of processes such as the influence of vertical shear and precipitation, superimposed upon gravity-current circulations.

2. Simulations of the 5 February 1978 system
a. Description of the model initialization

Our initial simulations are based on observations taken by Doppler radars of a cold front in central California on 5 February 1978. The details of the observations can be found in Carbone (1982, 1983) and Parsons et al. (1987). As previously mentioned, the system contained updrafts locally in excess of 20 m s⁻¹ despite the lack of appreciable potential instability in the warm sector. The horizontal scale of the updraft in this system was ~1 km. In order to accurately simulate the dynamics of an updraft and precipitation system on this scale, we employed a two-dimensional version of the Klemp and Wilhelmson (1978) cloud model. The horizontal axis was oriented perpendicular to the surface front (a nearly ENE–WSW alignment). The grid spacing was 250 m in the horizontal and vertical directions over a domain of 10 km in the vertical and 64 km in the horizontal. This grid spacing was chosen to facilitate comparison to the wind data in the observational study, which used a 300-m grid in both the horizontal and vertical directions.

The model used the complete set of compressible equations, Kessler-type microphysics for the cloud and rainwater fields, a radiative upper boundary condition, and a subgrid turbulent energy equation. The lateral boundaries were open with a free-slip lower boundary. The two-dimensional assumption is reasonable since observational evidence suggests that the frontal updraft is quasi–two-dimensional. For example, after finding a standard deviation in the vertical motion to be only 3 m s⁻¹ and only weakly dependent upon position, Carbone (1982) states “the quasi–two-dimensional characterization of the updraft is well founded.” Our simulations also did not include an ice phase. Microphysical modeling of the precipitation process in an intense frontal updraft (Rutledge and Hobbs 1984) suggests that the heavy precipitation associated with the updraft can result from a warm rain process.

The front was initialized relying primarily upon the previously discussed finding that the structure of the leading edge of mature cold fronts is similar to a gravity current. In laboratory studies gravity currents are typically formed from flow out of a deep reservoir of denser fluid. Numerical simulations of gravity-current flows are often initialized with either a cold reservoir (e.g., Rotunno et al. 1988) or a descending core of cold air (e.g., Mitchell and Hovemarle 1977). In this model we used a deep cold reservoir on the left-hand side of the domain (Fig. 1). The parameters needed to define the characteristics of cold reservoir are the thermal and humidity differences between the two air masses and
accurate portrayal of the observed front. The important problems of formation of a gravity-current front in a rotating atmosphere (Emanuel 1985; Gall et al. 1987) are not addressed in this study.

b. Control simulation

The flow vectors relative to the moving surface front and the accompanying precipitation fields for the simulation are shown at 1-hour intervals in Fig. 4. The system is dominated by warm-sector air in the lower levels approaching the surface front and rising rapidly just ahead of the leading edge of the cold air mass. The updraft is shallow, diverging substantially between 2 and 6 km in height into a branch that flows rearward over the cold air mass and an overturning branch that flows back into the warm sector. Within the cold air mass there is a weak relative circulation with air flowing toward the surface front above ~1 km and away from the front in the lowest 500–750 m. Heavy rainfall in the simulations, illustrated by the reflectivity pattern (Fig. 4) derived from the model rainwater content, is generated by the intense updraft resulting in an overhang of precipitation in the warm sector and a generally shallow area of heavy rainfall located just behind the surface front. The area of heavy reflectivity is narrow with the 55-dBZ contour typically less than 1 km wide at the surface.

The simulated flow in cross sections (Fig. 4) is relatively steady state. The variation of the maximum vertical motion with time is shown in Fig. 5. After the first hour, the maximum vertical motions ranged between 11.5 and 15.7 m s$^{-1}$. The time series confirms that the updrafts were relatively steady state, but it also reveals an oscillatory trend with a period of order 600–900 s and an overall slight decrease in updraft intensity.

![Fig. 1. A schematic of the initial conditions for the control simulation. The curve at the right represents the initial wind profile derived from a sounding taken ahead of the cold front at 1000 PST 5 February. This wind profile is in a coordinate system stationary to the simulated cold front. The variables $T_i$ and $q_{v_i}$ are the vertical profiles of temperature and water vapor mixing, respectively, obtained from a prefrontal sounding at 1000 PST 5 February 1978. The dotted line represents the initial location of the cold air mass while the shading and arrows indicate the subsequent movement of the cold air. The model boundary conditions are also indicated.](image1)

![Fig. 2. Wind hodograph representing prefrontal conditions from a sounding taken at 1000 PST 5 February 1978. The height of the wind measurements are indicated on the plot. The winds are in a frame of reference moving with the observed front and in a coordinate system rotated so that the $u'$ axis is perpendicular to the front. Note that differences between the wind profiles in Figs. 1 and 2 are due to small differences between the movement of the simulated and observed fronts. (Adapted from Carbone 1982.)](image2)
with time. The magnitude of the average updraft
(\(\sim 13.5 \text{ m s}^{-1}\)) is large, especially if one considers that
the strongest updraft occurs at a height of just above
2 km (Fig. 6).

Vertical cross sections of perturbation fields of pres-
sure and buoyancy for the simulated front at 3.5 h are
shown in Figs. 7a and 7b. These cross sections show
high pressure and negative buoyancy associated with
the cold air mass with slight warming and reduced
pressure evident above the cold air. The high-pressure
area extends into the warm-sector inflow ahead of the
front. Since the airflow and pressure distribution are
relatively steady, it is evident that as the warm-sector
air in the lowest level flows toward the front (see tra-
jectory in Fig. 7a), it stagnates at the leading edge of
the cold air and is associated with a pronounced pres-
sure rise of nearly 1.5 mb. Since the pressure rise occurs
in the relatively uniform buoyancy field ahead of the
front, the increase can be thought of as a dynamic pres-
sure change induced by changes in the wind field and
is not the result of vertical distributions of heating and
cooling. If the pressure rise ahead of the cold front is
dynamic and due to the stagnation of the warm-sector
flow at the leading edge of the front, the magnitude
should be approximately \(0.5U^2\) where \(U\) is the relative
inflow velocity. In this simulation \(U\) is \(\sim 17 \text{ m s}^{-1}\)
in the lowest \(\sim 500 \text{ m} (\text{Fig. 1})\) equating to a pressure rise
of \(\sim 1.4 \text{ mb}\), which is close to the value observed in
the simulation.

The dynamic pressure rise just ahead of the cold air
mass is important in terms of the vertical accelerations
associated with the intense frontal updraft. For a com-
pressible and nonrotating fluid an equation for vertical
acceleration (\(dw/dt\)) can be written as

\[
\frac{dw}{dt} = g\beta - c_p \frac{\partial T}{\partial z} + m_3,
\]

(1)

where \(w\) is the vertical motion and the symbols \(c_p\), \(\beta\),
and \(g\) represent the specific heat of dry air at constant
pressure, buoyancy, and the acceleration due to gravity,
respectively. The virtual potential temperature is given
by \(\theta_v\), and the pressure is expressed in terms of a non-
dimensional Exner function \(\pi = (p/p_0)R/\theta_v\), where
the subscript zero denotes a reference value and \(R\) is the
gas constant for dry air. The term \(m_3\) represents the
effects of mixing. The primary terms of interest are the
buoyancy and vertical pressure gradient. The buoyancy
term is usually the term driving cumulus convection
in the atmosphere, while the vertical pressure gradient
is crucial only in certain situations such as lifting air
parcels to their level of free convection, decelerating a
strong downdraft as it reaches the surface, and main-
taining steady, supercell convection.

In the present simulations, the rapid decrease in per-
turbation pressure with height ahead of and in the vic-
inity of the leading edge of the front (Fig. 7a) indicates
a strong upward-directed pressure force. Realizing that

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**Fig. 3.** (a) Skew T-log\(P\) plot of the sounding taken at 1000 PST 5 February 1978. The sounding, corresponding to the hodograph in
Fig. 2, represents the conditions ahead of the front. The data has been interpolated into the vertical levels used in the model. (b) Skew T-
log\(P\) plot of the prefrontal (1000 PST) and postfrontal (1300 PST) temperature soundings. Shaded area represents the temperature of lifted
parcels using a variety of assumptions concerning parcel and microphysical characteristics. The case is characterized by very small values
of the buoyancy. (Adapted from Carbone 1982).
the decrease of pressure with height approaches 1 mb km\(^{-1}\), one can estimate the pressure force to \(\sim 0.1\) m s\(^{-2}\) near the leading edge of the front (Fig. 7a). In contrast, the positive buoyancy in the updraft ranges from negligible to only slightly positive (Fig. 7b). Through examination of the magnitudes of buoyancy and vertical pressure forces in the vicinity of the updraft at various times we found that the buoyancy was only 4\% to 16\% of the upward pressure force. Thus, it is a combination of slight positive buoyancy and an intense vertical pressure gradient that results in the strong updraft. The shallowness of the updraft is consistent with both a downward-directed pressure force above \(\sim 2\) km (Fig. 7a) and the updraft becoming negatively buoyant above approximately 1.5 km (Fig. 7b). Although an upward-directed pressure gradient also exists within the cold air mass (Fig. 7a), there is little upward acceleration of the flow in this region, since the upward pressure force is balanced by negative buoyancy (Fig. 7b).

The general characteristics of the buoyancy field associated with the updraft (Fig. 7b) are consistent with parcel temperatures obtained from sounding data that shows some potential for a marginally unstable updraft below 2 km in height but an increasingly negatively buoyant plume if the parcel lifting extends above that height (Fig. 3b). The positive buoyancy in the middle levels to the rear of the leading edge of the cold pool (Fig. 7b) is also consistent with the differences between the prefrontal and postfrontal soundings in Fig. 3b, which shows a layer of relative warming between 2.1 and 6.1 km. The sounding data indicate relative cooling above that height, while the buoyancy field in the sim-
Fig. 7. (a) A vertical cross section of relative wind and perturbation pressure at 3.5 h into the simulation. The contour interval is 0.25 mb with pressure deviations in excess of 0.5 mb stippled. The solid line denotes a time-dependent trajectory through the system. (b) A vertical cross section of relative wind and buoyancy at 3.5 h into the simulation. The contour interval is $3.0 \times 10^{-4}$ with values greater than $3.0 \times 10^{-4}^{-4}$ shaded and less than $-6.0 \times 10^{-4}$ stippled.

Simulations (Fig. 7b) show the cooling occurring above 5 km.

We feel that the simulation represents a realistic depiction of the observed system. For example, a comparison between the simulated (Fig. 4) and observed (Fig. 8) airflow reveals that the kinematics in the two instances are similar, as both the simulated and observed flows are dominated by airflow from the lower levels of the warm sector rising rapidly at the leading edge of the front. As expected given the similarity between the observed and simulated flows, the general patterns of radar reflectivity (Figs. 4 and 8) are also similar with a narrow area of high reflectivity located just behind the leading edge of the surface cold front. The vertical profile of the average maximum updraft as derived from Carbone’s (1982) multiple Doppler radar measurements (Fig. 6) also closely agree with the profiles obtained from the model simulations, with the maximum occurring near 2 km.

The fields of buoyancy and pressure in the simulation (Fig. 7a and 7b) are somewhat similar to the pressure and buoyancy fields obtained by Parsons et al. (1987) from substitution of Doppler radar–derived winds into the equations of motion (Figs. 9a and 9b). For example, the strong upward pressure force, the lower pressure above the cold air mass, the low-level negative buoyancy associated with the cold air mass, the tendency for the densest air to be near the heavy precipitation region behind the front, and warming in a subsiding flow above the front are present in both the simulations (Figs. 7a and 7b) and the thermodynamically derived fields (Figs. 9a and 9b). While there are some quantitative differences between the retrieved and simulated thermodynamic fields (i.e., the low-pressure deficit above the cold air is far larger in the observations), the overall pressure distribution and physical interpretation of the flow dynamics are essentially the same. For example, the statement by Parsons et al. concerning the dynamics of the updraft based on the retrieved thermodynamic fields is directly applicable to the simulations. According to their study, “As the warm-sector flow collides with the cold air mass, a stagnation high pressure develops at the leading edge of the front. This pressure feature is at least partly nonhydrostatic, as evidenced by the intense vertical accelerations. The updraft is related to the high-pressure area decreasing with height with an upward directed pressure force.” The agreement afforded by this unique comparison between the observed and simulated thermodynamic fields both validates the use of the retrieval technique to gain dynamical insight into observed flows and adds further confidence to our ability to simulate this system.

Thus far, we have treated the comparison between the simulated and observed flows with a strong degree
of optimism. However, we realize that there are some differences between the two flows. For example, a comparison between the two average maximum updrafts suggests that the model may underpredict the observed maximum by 5% to 10% (Fig. 6). However, this difference is within the variations in space and time for the observed flow (Carbone 1982) and within the range of error expected from multiple Doppler analysis. The model-estimated maximum radar reflectivity is \( \sim 3-4 \) dBZ higher than the observed values. It is not known whether this discrepancy is the result of actual differences in precipitation mass between the observed and simulated fields, since the radar does not measure precipitation mass directly. Since the model microphysics does not include an ice phase, one should not expect an exact agreement.

Another difference between the observed and simulated flows (Figs. 4 and 8) is that the observed field seems to have more finescale structure than the simulations. A possible source of this finescale structure is errors in the derivation of three-dimensional winds from the Doppler radar observations. In addition, Carbone (1982) discussed the possibility of shear-induced waves behind the front. These waves could be lacking in the simulated flow due to a higher Richardson number, since the simulated flow is two-dimensional and does not include the contribution to the vertical shear from the component of the horizontal wind parallel to the front. Another difference between these flows is that the wind above \( \sim 5 \) km in the simulations has a larger easterly component than was present in the observations. In the subsequent section we will present evidence that shows that the flow at this upper level does not seem to significantly impact the strength of the frontal updraft. Another difference between the two flows is that the observed flow within the cold air mass does not reveal a return flow away from the surface front in the lower layers. However, this difference is simply an artifact of the multiple Doppler analysis, as one of the radars was located on elevated terrain, making reliable multiple Doppler winds unavailable below approximately 750 m for the distances from the radar used in this study. The postfrontal hodograph, single Doppler measurements, and schematics presented by Carbone (1982) all reveal a return flow in the lowest \( \sim 500 \) m of the cold air as is evident in the simulations.

c. Simulations with various wind profiles

The previously discussed results show that the intense upward acceleration associated with the shallow frontal updraft is driven primarily by a dynamic pressure force. An initialization with the identical thermodynamic sounding but with constant wind was carried out as a crude first step to isolate the effects of vertical shear on the dynamic pressure forces and on the corresponding updraft strength. The time series of maximum vertical motion for the control and constant wind simulations are shown in Fig. 5. While the updraft magnitudes remained quasi–steady state in both cases, the updraft in the no-shear case is substantially smaller throughout the simulation. The plot of the maximum vertical motion as a function of height for the constant wind case (Fig. 10) reveals a far shallower updraft maximum than we found in either the control simulation or the observed system (Fig. 6).

The relative winds and the reflectivity field for this no-shear simulation further illustrate a shallower and weaker ascent (Fig. 11). The maximum vertical motions (6 m s\(^{-1}\) to 14 m s\(^{-1}\)), the height of the maximum updraft (0.75 km to 2.0 km), and the rainwater content (\( \sim 0.5 \) g kg\(^{-1}\) to \( \sim 5 \) g kg\(^{-1}\)) are all substantially less in the simulation without vertical shear. In addition to the precipitation contents being far lower in the no-shear case, the precipitation maximum at the surface is located \( \sim 7 \) km to the rear of the frontal wind shift,
while the maximum in the ambient shear (control) case takes place ~2 km behind the frontal wind shift. The difference in the location of the precipitation in the two cases is the result of the greater tilt of the updraft in the constant wind case, while the pronounced, nearly order-of-magnitude difference in the precipitation rates is due to the deeper and more intense updraft in the control case.

One may be tempted to conclude that the stronger updraft in the ambient wind case is simply the result of greater surface convergence. A comparison between the low-level convergence fields in the two cases (Fig. 12), however, reveals that the maximum low-level convergence was greater in the no-shear case, but the convergence at the leading edge of the cold front was deeper in the ambient shear case, resulting in a deeper and more vigorous updraft. This result is identical to the Rotunno et al. (1988) cold-pool simulations with no shear and “optimal” shear (see their Figs. 20a,b) in which they also found the strongest low-level convergence in the constant wind case but a more favorable vertical distribution of convergence in the optimal case. (In their paper and throughout this study, we will refer to the value of vertical shear that results in the strongest updraft at the leading edge of the cold pool as the optimal vertical shear.)

In order to investigate what characteristics of the ambient wind profile produce the dramatically stronger updraft, we removed the vertical shear in various layers. While all of these simulations will not be discussed, we will show two different simulations to illustrate our findings. In one simulation, the ambient shear below 2.5 km was retained but there was no shear above 2.5 km. In the other simulation there was no shear below 2.5 km but the ambient profile was prescribed above that level. The flow vectors and precipitation for these two simulations are shown in Fig. 13. The relative flow and precipitation pattern in the simulation with the ambient shear in the lowest 2.5 km (Fig. 13a) is similar
to the control case (Fig. 4) except for slightly less rainfall than in the control and a flow above \( \sim 5 \) km that is a closer match of the observations (Fig. 8) than the control case. A possible explanation for the latter result is that the sounding used to initialize the control simulation was not from an undisturbed environment but was influenced by the frontal circulations, which modified the flow aloft.

In contrast, the simulation without shear in the lowest 2.5 km produced a weak updraft (Fig. 13b) similar to that observed in the constant wind simulation (Fig. 11). The updraft in this simulation was substantially lower in depth and magnitude than was observed in any of the simulations that included ambient vertical shear

![Figure 13](image)

**Fig. 13.** (a) Vertical cross section of relative wind and derived radar reflectivity for the simulation at 2 h with constant wind above 2.5 km. The radar reflectivity contours are shaded at the 45- and 55-dBZ levels. (b) Vertical cross section of relative wind and derived radar reflectivity for the simulation at 2 h with constant wind below 2.5 km. The radar reflectivity contour is 35 dBZ.

3. Idealized simulations

In order to test the hypothesis that the strong updrafts were due to the interaction of the low-level vertical shear and the cold front, we devised a number of simplified numerical experiments. The experimental design is shown in Fig. 14. The initial idealized simulations were “dry” as the influence of precipitation was eliminated by setting the water vapor values in the model to zero. In order to preserve the near-neutral ascent found in both the observations and the control simulation, the lapse rate in the lowest 4 km was set to be dry adiabatic. This neutral zone was capped by a stable layer with an increase in the potential temperature with height equal to \( 4^\circ \)C km\(^{-1} \) (Fig. 14). The domain for these idealized simulations encompassed 128 km in the horizontal and 10 km in the vertical with a 500-m and 250-m grid spacing in the horizontal and vertical directions, respectively. (The larger horizontal grid spacing was used in order to conserve computational resources.) An initial cold pool of variable depth and magnitude was placed in the left-hand portion of the domain. In order to isolate the effects of low-level vertical shear, a wind profile similar to that used in Thorpe et al. (1982) and Rotunno et al. (1988) was prescribed with constant linear shear below 2.5 km and constant wind aloft.

### a. Variation of the low-level vertical shear

The magnitude of the vertical shear in the lowest 2.5 km was varied in an effort to isolate the impact of vertical shear upon the frontal updraft. In this experiment the initial depth of the cold pool was 2 km. The cold air flowed out of this reservoir in a gravity-current flow similar to but shallower than that produced in the earlier experiment (e.g., Fig. 7b). The plot of maximum updraft at 2 h into the simulations for various vertical shears (Fig. 15) shows that the strongest updraft of \( \sim 8.5 \) m s\(^{-1} \) occurred with a vertical shear of \( 20 \) m s\(^{-1}/\) 2.5 km. In contrast, the updraft for the constant wind case was only \( 4.5 \) m s\(^{-1} \). The height of the maximum updraft also had a direct dependence upon the vertical shear, with the strongest updraft located at a height of 2.25 km for the optimal case and near 1.0 km for the constant wind simulation (Fig. 15). The combined impact of the dependence of the updraft magnitude

![Figure 14](image)

**Fig. 14.** As in Fig. 1 but a schematic of the initial conditions for the idealized experiments. The depth of the cold pool, the magnitude of the vertical shear, and the depth of the neutral layer were all varied in subsequent experiments.
and height upon vertical shear produces a dramatic difference in the nature of the updrafts evidenced in the airflow for simulations with low, optimal, and high shears (Fig. 16). The low-shear simulation shows a shallow updraft sloping over the cold pool (Fig. 16a), while the near-optimal case reveals an intense, nearly vertical updraft (Fig. 16b). The idealized simulation with higher than optimal shear contains an updraft that leans downshear into the warm air mass (Fig. 16c). These simulations clearly show the strong dependence of the updraft strength and character at the leading edge of gravity-current fronts on vertical shear. Since the initial cold-pool depth of 2 km in these idealized experiments was well less than the initial 5-km cold-pool depth needed to replicate the observed system, this series of simulations also demonstrate the ability of modest cold pools in the presence of strong vertical shears to produce intense updrafts.

The dependence of the updraft slope upon vertical shear is consistent with the previously discussed theoretical work of Rotunno et al. (1988). In their study, they proposed that the strongest updraft occurred with an “optimal” vertical shear and that this optimal state took place when the horizontal vorticity generated by the horizontal gradient of buoyancy across the front was balanced by the horizontal vorticity associated with the ambient vertical shear. This balance can be expressed mathematically (Rotunno et al. 1988) as

$$\Delta u^2 = 2g \int_0^H (-\beta) dz = c^2,$$  

where $\Delta u$ is the horizontal wind change with height through the vertical layer interacting with the cold pool, $\beta$ is the buoyancy term taken in an integral over the depth of the negatively buoyant area, $H$ is the depth of the negatively buoyant air mass, $g$ is the acceleration due to gravity, and $c$ is the speed of the cold pool. When the ratio of $\Delta u/c$ is approximately 1, Rotunno et al. argue, the balance is optimal and a deep and intense updraft results.

In order to test whether the strongest updraft in our idealized simulations occurred near the optimal vertical shear predicted by the Rotunno et al. (1988) study, we calculated the rhs of (2) for the model simulation with the optimal vertical shear by estimating the average integrated buoyancy at various locations well behind the leading edge of the front. Using this technique, we estimated $c$ to be 21.8 m s$^{-1}$. Due to spatial variations in the estimated integrated buoyancy we feel that the variation in this calculation is $\sim 2$ m s$^{-1}$. Thus, from (2) we predict that the strongest updraft should occur for a shear between 18.8 and 23.8 m s$^{-1}$, in reasonable agreement with the optimal vertical shear of 20 m s$^{-1}$ found in Fig. 15. A comparison of the magnitude of the updrafts in the theory and the simulations is not possible, since the theory does not predict updraft magnitude.

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**Fig. 15.** A plot of maximum updraft versus low-level vertical shear for a series of simulations with an initial cold-pool depth of 2 km (solid line) and 4 km (dashed line). The height of the maximum updraft is labeled alongside the curves. The predicted value of the maximum updraft is indicated by an arrow.

**Fig. 16.** Relative airflows for idealized simulations with an initial cold-pool depth of 2 km. The updraft magnitude is contoured and shaded at the 2.5 and 5 m s$^{-1}$ levels. Shown in these figures are (a) constant wind, (b) optimal shear (20 m s$^{-1}$/2.5 km), and (c) high-shear simulations (32 m s$^{-1}$/2.5 km).
The value of $c$ in (2) represents the theoretical speed of an inviscid gravity current (e.g., Benjamin 1968). The speed of the cold air mass in this idealized simulation is $\sim 15 \text{ m s}^{-1}$, which is well less than the predicted value of $c$. The observation that the frontal movement is slower than this theoretical value is expected, given the well-known finding that gravity currents in the atmosphere and laboratory move typically at a speed approximately $3/4$ of this theoretically predicted speed (e.g., Simpson 1969). Benjamin (1968) also argued theoretically that a more appropriate estimation of the density current speed could be obtained by replacing the constant 2 in (2) by 1.1. As Rotunno et al. noted, this well-known difference between the observed and theoretical speed of a gravity current should be kept in mind when one attempts to compare the Rotunno et al. theory with observations, since the observed movement of the cold air mass will consistently underpredict the amount of vertical shear required for an optimal balance. While the actual movement of the cold air is not a reliable predictor, our calculation suggests that estimation of the optimal shear through substituting the integrated buoyancy into (2) does result in reasonable agreement.

b. Change in the cold-pool depth

Another idealized simulation was undertaken with the initial depth of the cold pool doubled from 2 km to 4 km. This initial depth is more relevant to the simulations of the 5 February system. The magnitude and height of the maximum updraft versus the vertical shear are shown in Fig. 15. In the simulations with deeper cold pools, the characteristics of updraft are again strongly dependent upon the ambient vertical shear. In the simulations with a deeper initial cold pool the maximum updraft for the constant wind case was only $\sim 5.9 \text{ m s}^{-1}$ at a height of 1.25 km, while the value for the case with optimal vertical shear was $\sim 15.1 \text{ m s}^{-1}$ at a height of 2.75 km.

The magnitude of the optimal vertical shear for the simulation with the deeper cold pool is within 10% of the value predicted by (2) (Fig. 15). Given the accuracy in our predicting the optimal state, the predicted value is again close to that found in the simulations. However, there are reasons to expect some discrepancy between the predicted value and the optimal values. For example, the Rotunno et al. study examined the balance between only two horizontal vorticity sources; the low-level shear in the inflow and the horizontal vorticity generated by buoyancy at the leading edge of the cold pool. Our initialization contained vertical shear within the cold air mass that results in a third source of horizontal vorticity. Since we found that the optimal shear in the simulations is reasonably close to that calculated from (2), we can conclude that the system is not very sensitive to this third source of horizontal vorticity.

The optimal shear for the deeper cold pool was approximately $32 \text{ m s}^{-1}$ in the lowest 2.5 km, which represents an approximately 60% increase over magnitude of the optimal shear in the idealized simulation with the shallower cold pool. The increase in optimal shear with cold-pool depth is expected from examination of (2). A comparison of the maximum updrafts in the two optimal cases reveals that the updraft in the case with the deeper initial cold pool is far larger than in the simulations with the shallower cold pool. This dramatic increase in updraft strength in the case with the deeper cold pool is due to two factors: 1) an increase in the magnitude of the updraft for a fixed value of the vertical shear that results from a deeper cold pool and 2) a larger optimal vertical shear. While our discussion has partly concentrated upon the optimal-shear cases, these simulations also indicate that strong and relatively deep updrafts can occur for a wide range of modest shear.

c. Variation in the depth of the neutral inflow

In the previously discussed observations, common features of the warm-sector environment were strong low-level vertical shear normal to the front and near-neutral stability to low-level parcel ascent. Another idealized experiment was conducted with the depth of this neutral layer varied, while the vertical shear and initial depth of the cold pool were held constant. The gradient of potential temperature with height above the neutral layer was $4 \text{ C km}^{-1}$. In these simulations, the initial depth of the cold pool was 2 km and the vertical shear was set at its optimal value of $20 \text{ m s}^{-1}$.

We found the magnitude and height of the maximum updraft in the optimal case to be dependent upon the depth of this neutral layer with the value of the maximum updraft varying by approximately a factor of two as the neutral layer depth is changed from 2 to 10 km (Fig. 17). The dependence of the updraft strength upon the depth of the neutral layer is weaker for a series of simulations with constant wind (Fig. 17). This behavior is easily understood. In the optimal case, the updraft is deep and low-level parcels are lifted into the stable layer where they are negatively buoyant and contribute to a deceleration of the updraft. The height at which this deceleration is realized is directly dependent upon the depth of the neutral layer. In the constant wind simulations, the lifting is shallow so that low-level parcels do not penetrate into the stable layer aloft and the depth of the neutral layer does not greater influence the updraft magnitude.

d. The effects of moisture

The previously presented simulations illustrate the potential for gravity currents to produce intense updrafts in environments with moderate to strong vertical shear in the cross-frontal direction and near-neutral
stability in the lower levels of the ambient environment. However, in order to demonstrate that this mechanism is relevant to observed fronts we repeated these experiments with moisture. A series of simulations were conducted with an idealized wind profile and the ambient thermodynamics of the 5 February system (Fig. 3a). The initial depth of the cold pool was 4 km and the wind profiles were represented by constant vertical shear below 2.5 km and constant wind aloft (i.e., as in Fig. 14).

The plot of the height and magnitude of the maximum updraft as a function of vertical shear (Fig. 18) reveals the same pattern as in the simulations without moisture, in that the depth and intensity of the updraft increases dramatically with vertical shear toward an optimal value. The optimal value is again reasonably close to that predicted by (2). At greater than optimal shears, the presence of precipitation complicates the situation as an unsteady mode of convection results, as precipitation from the downshear tilted updraft falls into and interrupts the inflow (Fig. 19). In these high-shear simulations the leaning of the cold pool downshear also creates potential instability as the cold air overlies the warm-sector flow (Fig. 19). The combination of the creation of potential instability and the interruption of the inflow by the precipitation creates a unstable flow unlike any airflow observed to date at the leading edge of cold fronts. It should be noted that dry simulations with deep cold pools and extreme values of the vertical shear also demonstrated some tendency for unsteady behavior. Our finding that the strongest updraft in a heavily precipitating environment has a nearly vertical orientation is somewhat surprising since one might expect that vertical updrafts would tend to be weakened by increased water loading. Instead, our simulations demonstrate that the stronger vertical pressure forces associated with the optimal state outweigh any increased negative buoyancy due to water loading.

4. Discussion of results

We have found that our simulations with the ambient vertical shear closely replicate the intense frontal updraft frequently observed near the leading edge of sharp cold fronts. The agreement between the observed and simulated front with an initialization with a deep reservoir of cold air provides further support (along with observational evidence described in the Introduction) for the idea that sharp surface fronts are well represented by an interaction of a gravity current with the ambient environment. Since the observed front is one of the more intense cases presented in the literature, further simulations of less active fronts are clearly needed.

We do, however, acknowledge that there are some distinct differences between the frontal flows and classical dry gravity currents. For example, as pointed out by Smith and Reeder (1988), the observed flow within the cold air mass in the 5 February case is reversed from that found in a dry gravity current. We found this flow reversal to occur in our moist simulations with near-optimal vertical shears (i.e., Fig. 4). While Carbone (1982) had hypothesized that the return flow

![Fig. 17. Plots of the maximum updraft versus depth of the neutral layer ahead of the front for the optimal shear (solid line) and constant wind (dashed line). The height of the maximum updraft is labeled along side of the curves.](image)

![Fig. 18. A plot of maximum updraft versus low-level vertical shear for a series of simulations with thermodynamics derived from a prefrontal sounding (Fig. 3a) and an initial cold-pool depth of 4 km. The higher than optimal shear values are not plotted due to their unsteady nature. The predicted optimal value is also shown.](image)
Reeder (1988) used the differences between observed fronts and classical gravity-current theory to discount the gravity-current model of cold-frontal circulations. Our simulations show that effects such as vertical shear and the precipitation loading due to heavy rainfall superimposed upon gravity-current circulations can account for some of the differences between the microscale structure of observed fronts and classical dry gravity currents.

While our simulations do support the hypothesis that the finescale structure of fronts is well approximated by gravity-current dynamics, they also show that the presence of a gravity current alone is not sufficient to develop deep, intense updrafts and an accompanying narrow band of heavy rainfall. In our simulations, we found that it is the interaction of a gravity current with the ambient vertical shear in the cross-frontal direction that results in very intense updrafts in environments nearly neutral to vertical ascent. Under these conditions, a narrow band of heavy rainfall is generated by strong, deep lifting of air that is nearly saturated. In environments of weak vertical shear, the lifting is shallower and sloped, which results in a weaker and broader area of precipitation located farther behind the leading edge of the front.

The intense vertical accelerations in these simulations are not caused by positive buoyancy but instead are associated with an intense upward-directed pressure force. Since rawinsonde measurements ahead of observed intense frontal updrafts and their accompanying narrow cold-frontal rainbands (e.g., Browning and Harrold 1970; Carbone 1982; Hobbs and Persson 1982; Parsons and Hobbs 1983b) also show negligible potential instability, we feel that there is now very strong support for the hypothesis that intense frontal updrafts and the narrow band of heavy precipitation are primarily a consequence of these vertical pressure forces at the leading edge of the cold air mass. The vertical pressure force in both our simulations and past observations was associated with a high pressure area in the lower levels that in turn was associated with intense convergence as the warm-sector inflow stagnated near the leading edge of the cold air mass. Hence, the pressure forces were primarily dynamic in that they were not associated with vertical distributions of buoyancy. The deeper and more intense updrafts in the simulations with vertical shear had a more favorable distribution of convergence and hence a stronger upward-directed pressure force.

The dependence of the simulated updrafts on vertical shear (i.e., the updraft increases with the addition of vertical shear until an optimal value is reached, after which the strength and depth of the vertical motion begins to decrease) is similar to the behavior of updrafts at gravity currents, as predicted by Rotunno et al. (1988). The change in the slope of the updraft with vertical shear (i.e., Fig. 16) is also consistent with the Rotunno et al. theory. The simulations in our study
quantify the Rotunno et al. mechanism and demonstrate its applicability to frontal systems. The Rotunno et al. theory was partly developed in response to numerical simulations of squall lines (which were later shown in Weisman et al. 1988) that found the slope and magnitude of the convective updrafts changed during the lifetime of the squall line system. This evolution was interpreted as result of a modification of the horizontal vorticity balance by a gradual increase in the strength of a convectively induced cold pool while the vertical shear in the lower levels ahead of the cold air remained relatively constant. The strong time dependence of the squall line cold pools is clearly absent in our simulations, and past observations of these frontal systems as a lack of strong evaporatively cooled downdrafts results in a relatively steady cold pool. As stated earlier, this result is to be expected since these shallow frontal updrafts occur in environments that are nearly saturated and with lapse rates that are nearly moist adiabatic.

The optimal vertical shear in the Rotunno et al. (1988) theory was derived from examination of the horizontal vorticity budget in the vicinity of the cold pool. Since the optimal state in our simulations is relatively well predicted by this theory, even those with moisture, we can conclude that the theory is generally applicable to this type of cold front. Lafore and Moncrieff (1989) cautioned against the application of this theory to deep convective systems due to the possible generation of horizontal vorticity by horizontal buoyancy gradients not associated with the leading edge of the cold pool. While we acknowledge that even our simple frontal simulations do contain other sources of horizontal vorticity, such as with the initial flow within the cold pool or the flow reversal driven by precipitation within the cold air, the general agreement between the optimal state predicted by (2) and that found in the simulation suggests that these other sources are secondary for these fronts. For example, the impact of this flow reversal on the balance can be illustrated in the idealized simulations with precipitation. In these simulations the optimal vertical shear is \( \sim 18 \text{ m s}^{-1} \). Substituting this vertical shear into (2) where the shear is squared results in a contribution of \( \sim 324 \text{ m}^2 \text{ s}^{-2} \) toward balancing the integrated buoyancy. In contrast, the magnitude of the vertical wind change associated with the flow reversal in the cold air is \( \sim 6 \text{ m s}^{-1} \) over the depth of the cold pool for a contribution toward the balance of only \( \sim 36 \text{ m}^2 \text{ s}^{-2} \). Hence, the contribution to the balance by the flow reversal is nearly an order of magnitude smaller than that associated with the vertical shear in the warm-sector flow. This result, while expected since the theory predicted the optimal state reasonably well, illustrates that for this system the primary terms within the balance are the buoyancy gradient associated with the leading edge of the front and the vertical shear in the ambient flow.

Weisman (1990) found specific squall line environ-

ments where deep convection became organized in a
such a way so that other sources of horizontal vorticity
needed to be added to the balance. In frontal cases with
limited potential instability and small buoyancy gra-
dients apart from those associated with the leading edge
of the front, it appears that these other sources of hori-
zontal vorticity are clearly secondary. Hence, these
frontal environments may provide a more direct test
of the Rotunno et al. (1988) squall line theory than
those systems containing deep convection. While the
details of the theory appear generally applicable to our
frontal simulations, it is not the details of the balance
that are the important but rather the fact that the in-
teraction of the leading edge of the cold pool with
the vertical shear provides an strong influence on the char-
acter of the updraft.

An alternative method of explaining the interaction
of cold pools with vertical shear has been proposed by
Thorpe et al. (1982). In their theory the vertical shear
is important in that it influences differences in move-
ment between the cold pool and that of the convective
cells above the cold pool. The optimal case is when
these two speeds are equal. In this point of view, there
is no importance whatsoever in whether there is any
vertical shear within the inflow layer that interacts with
the cold air mass. In order to test this hypothesis, we
elevated the vertical shear layer to near the top of the
cold pool, while keeping the flow in the lower layers
constant (Fig. 20). While these simulations are not a
fair test of the Thorpe et al. hypothesis since the case
is near neutral and does not have convective cells above
the cold pool, the simulation with elevated vertical
shear does show the impact of an elevated shear profile
upon the lifting. In this simulation (Fig. 20) the updraft
is shallow and weak in a manner similar to the constant
wind simulation, which does not support the Thorpe
et al. hypothesis but is instead indirect support for the
hypothesis that it is the interaction between the vertical
shear and the cold air mass that strongly impacts the
ascent at the leading edge of the cold air.

5. Conclusions

The simulations in this study, together with past ob-
servational work, show that strong updrafts can occur
at the leading edge of gravity currents in the absence
of potential instability in the ambient environment. The
conditions that lead to these intense updrafts in-
clude 1) strong cold pools, as evidenced by the large
values of integrated negative buoyancy; 2) appreciable
low-level vertical shear; and 3) stability ahead of the
front that is nearly neutral to low-level parcel ascent.
Past observational studies of narrow cold-frontal rain-
bands (e.g., Browning and Harrold 1970, 1971;
Browning 1971, 1974; Browning and Pardoe 1973;
Carbone 1982; Hobbs and Persson 1982; Parsons and
Hobbs 1983a, b) indicate that the environment asso-
ciated with strong, small-scale updrafts at the leading
edge of the front tends to include ample low-level shear

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associated with a low-level jet, a sharp gravity-current-like cold front, and conditions ahead of the front that are nearly moist adiabatic and saturated. Hence, it seems likely that the mechanism discussed in this study, in which the cold pool interacts with the low-level vertical shear, plays a major role in causing strong updrafts just ahead of these sharp fronts in the atmosphere. In environments that are nearly saturated, the strong updraft that results from this interaction produces a narrow band of heavy rainfall. At low values of vertical shear the updraft is sloped, which in nearby saturated environments results in a relatively weak broad band of rainfall located well behind the surface front.

The findings presented here have a number of implications. First, they illustrate that when air is sharply lifted by an advancing pool of cold air, vertical pressure forces are important and may even dominate the dynamics. Thus, one is cautioned about simply applying ideas derived from parcel theory to lifting by cold pools. Second, it implies that our ideas concerning the lifting of air by sharp fronts are more complex than either the large-scale view that suggests that the strong updraft is the result of frictional convergence (i.e., friction turning the wind and increasing the component of the flow toward the front) (Keyser and Anthes 1982; Knight and Hobbs 1988) or the microscale view that simply having a gravity-current front will result in a strong updraft and a narrow cold-frontal rainband (Shapiro et al. 1985; Parsons et al. 1987). The findings portrayed in this study suggest low-level shear, in the nearly neutral environments described previously, provides a strong influence on the strength of updrafts at fronts. In this framework it is the vertical distribution of convergence (and hence dynamic pressure) that produces the strong updrafts with frictional turning important not for the increasing surface convergence but for the degree of vertical shear it implies.

Finally, we state that the simulations presented here also provide further support for using gravity-current theory as a conceptual model for sharp atmospheric cold fronts. The study does not represent an attempt to develop a theory for the formation of gravity-current fronts from large-scale processes. Thus, the study does not address current views of frontogenesis, but instead relies upon observations that demonstrate the existence of these fronts to initialize the simulations and then draw conclusions from these simulations. It should be noted that the existence of a gravity-current cold front does not automatically invalidate our classical ideas concerning fronts. For example, the movement of a such a front should be adequately represented in mesoscale models since the movement of a gravity current is based upon the hydrostatic pressure difference between the warm and cold air masses. However, our study does demonstrate that the airflows and accompanying weather at the leading edge of the front are sometimes explained by gravity-current dynamics. In addition, the presence of certain flow characteristics, such as the mesoscale extent of the flow reversal in the cold air due to the heavy rainfall, suggests that the processes occurring at the leading edge of the front have an impact on the mesoscale and may need to be represented or parameterized in order to accurately represent some mesoscale frontal flows.

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