Microphysical Processes Associated with Intense Frontal Rainbands
and the Effect of Evaporation and Melting on Frontal Dynamics

MARY C. BARTH AND DAVID B. PARSONS

National Center for Atmospheric Research, * Boulder, Colorado

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ABSTRACT

Previous studies have shown that a surface cold front often coincides with a heavy band of precipitation commonly designated as a narrow cold-frontal rainband. The maximum rainfall rate within this band can exceed 100–200 mm h⁻¹. This study uses a nonhydrostatic two-dimensional cloud model with ice microphysics to investigate the precipitation processes within this type of rainband. Despite the relatively simple initialization and two-dimensionality, many aspects of these storms were well simulated. In these simulations, the intense but shallow updrafts produced large amounts of cloud water that were transformed primarily into rain and graupel within the zone of heavy precipitation and, to a lesser extent, into snow. The graupel and snow produced a zone of trailing stratiform precipitation. While the heavy rainfall could be represented in a warm rain model of the storm, an ice phase was needed in order to replicate the stratiform precipitation. Feedbacks of microphysical processes upon the dynamics of the flow were investigated. Sublimation and melting of frozen hydrometeors produced a pronounced cooling within the cold air mass, which slowly increased the depth and intensity of the cold air mass. This diabatic cooling within the cold air could potentially play a role in maintaining or even intensifying the circulations that lead to these rainbands. Previous studies of these types of fronts have instead concentrated on the role of melting in maintaining these structures through producing a stable layer across the cold air interface that could inhibit mixing.

1. Introduction

A narrow band of heavy rainfall that is a few kilometers wide frequently accompanies the passage of surface cold fronts, particularly in maritime locations in the middle latitudes. This band of heavy rainfall is often referred to as line convection (Browning and Harrold 1970) or a narrow cold-frontal rainband (NCFR) (Houze et al. 1976). The maximum surface rainfall rate in NCFRs can exceed 100–200 mm h⁻¹ for periods of several minutes. Radar and aircraft studies of NCFR events (e.g., Browning and Pardoe 1973; Hobbs and Biswas 1979; James and Browning 1979; Matejka et al. 1980; Carbone 1982; Hobbs and Persson 1982; Parsons and Hobbs 1983a,b; Bond and Fleagle 1985) have revealed that these heavy rainfall rates are associated with intense vertical motions that are surprisingly shallow in vertical extent, with the strongest updraft typically found between 1 and 2 km in height. The potential instability in the warm sector air flowing into the NCFR is often near neutral. Instead, this strong ascent is highly nonhydrostatic as the updrafts are driven by a vertical pressure gradient (Parsons et al. 1987; Parsons 1992). The intense low-level horizontal convergence accompanying these nonhydrostatic updrafts correspond to near discontinuities in the temperature and surface winds present at the leading edge of a cold front. Carbone (1982) and Parsons et al. (1987) presented evidence to show that the circulations at the leading edge of a cold front that lead to the strong vertical motions in an NCFR resemble the forced ascent at the leading edge of a density current. Rotunno et al. (1988) proposed, and Parsons (1992) showed, that the interaction of a density current with the vertical shear in the ambient flow for this case may account for the stronger ascent at the leading edge of the front. Steady-state analytical models based on the assumption of no mixing between the air masses also have proven useful in understanding the relationship between NCFRs and density current dynamics (e.g., Moncrieff and So 1989).

The frontal system studied in Carbone (1982, 1983) and Parsons et al. (1987) was particularly intense with an updraft locally in excess of 20 m s⁻¹ and surface rainfall rates in excess of 100 mm h⁻¹ for over 4 min. The airflow associated with this front and the accompanying NCFR were well observed since measurements were taken by a triple-Doppler-radar network as this
frontal system passed through the central valley of California. Rutledge (1989) used these airflows to initiate a three-dimensional kinematic cloud model to study the production of precipitation in this heavy rainfall event. These kinematic simulations were used to infer that the heavy rainfall was generated through the production of graupel by riming and through coalescence and collection processes at temperatures above 0°C.

The Rutledge (1989) study made another inference from these simulations in that they found that the diabatic effects were important in maintaining the thermal differences across the cold front. Both the Carbone (1982) and Rutledge studies stressed that the melting of ice phase particles was important through enhancing stability and therefore decreasing mixing across the cold air interface in the stratiform precipitation behind the leading edge of the front, while Parsons et al. (1987) found evidence for diabatic cooling within the cold air mass. These hypotheses have implications for understanding the longevity of these systems.

While the Rutledge (1989) investigation and a similar study of a previous case (Rutledge and Hobbs 1984) does provide insight into the microphysical processes within NCFR systems and may provide estimates of diabatic forcing on the cold front, no temporal evolution of the airflow due to diabatic forcing can be addressed in a model with a steady-state airflow. Investigation of this feedback is needed since the formation of these density current fronts and their maintenance in a rotating atmosphere is currently unresolved. The diagnosed diabatic forcing determined by Rutledge (1989) is likely to be in error for this case since the thermodynamic energy equation used by Rutledge employs the hydrostatic assumption while the strong shallow vertical motions observed in this case have been shown to be highly nonhydrostatic. In this study we use a two-dimensional version of the Klemp–Wilhelmson (1978) model with a mixed-phase microphysical parameterization to study the microphysical processes that take place within the storm first studied by Carbone (1982) and the feedback of these processes upon the dynamics of the rainband and accompanying frontal system. The two-dimensional model is an inherent limitation of our approach, since the front contained three-dimensional circulations (Carbone 1982). However, the two-dimensional approach may be generally valid, as supported by statements in Carbone [(1982), “the prefrontal updraft is purely two-dimensional”] and in Rutledge [(1989), “the essential components of the precipitation process in this NCFR will be shown to be essentially two-dimensional”].

2. Model description and experimental design

The numerical model used in this study is a two-dimensional version of the Klemp and Wilhelmson (1978) cloud model that has been modified to include ice microphysics. This model uses the complete set of compressible equations and parameterized microphysics to predict the horizontal and vertical wind fields, potential temperature, pressure, water vapor, cloud water, rain, ice, snow, and graupel. The subgrid turbulence parameterization follows Deardorff’s method and is outlined in Klemp and Wilhelmson (1978). The vertical velocity is set to zero at the upper and lower boundaries. The lower boundary is free slip, and the lateral boundaries are open. The vertical domain of the model is extended to aid in deterring gravity waves from interfering with the storm dynamics. The model domain is 100 km horizontally by 15 km vertically, with a horizontal grid spacing of 0.5 km and a vertical grid spacing of 0.25 km. To keep the simulated NCFR centered in the model domain, the model domain followed the general movement of the horizontal location of the maximum updraft. For the simulations discussed below, the time step was 5 s.

The microphysics parameterization includes five hydrometeor reservoirs (cloud water, rain, ice, snow, and graupel). The cloud water is described with a Khrgian–Mazin size distribution (Rogers and Yau 1989); ice is assumed to be monodispersed; and rain, snow, and graupel are described with an exponential decay size distribution (Marshall and Palmer 1948; Gunn and Marshall 1958). The microphysical processes are parameterized using a modified version of the Rutledge and Hobbs (1983, 1984) scheme (which uses a modified Kessler parameterization). The ice enhancement routine is parameterized as described in Rutledge (1989). The method for calculating collection processes between rain, snow, and graupel has been modified according to Mizuno (1990).

The narrow cold-frontal rainband studied was observed on 5 February 1978 near Sacramento, California (Carbone 1982). This particular storm was unique in that it contained strong updrafts (~20 m s^-1) in a shallow vertical depth (~5 km). Because of these unique characteristics and the availability of detailed observations, this system has been investigated in detail as noted in the introduction.

For the simulations undertaken for this study, the temperature, water vapor content, and horizontal wind were initialized using the Sacramento sounding (Fig. 1) from 2 h prior to the passage of the storm, which indicated an environment close to neutrally stable to moist ascent. To initiate frontal lifting, a deep cold reservoir on the left side of the model domain was used to replicate a density current. The initial depth of the reservoir was specified at 5 km in the control simulation but, for sensitivity studies, was set at different levels. This cold reservoir was six degrees colder than the environment and was drier (67% of the environmental relative humidity). Accounting for the temperature and humidity perturbations, the relative humidity at the lowest model level in the cold pool was 91.6%, similar to the 91% at the lowest model level outside the cold pool. In these simulations, the cold air flows out of this
reservoir and produces a density current that is similar to the observed flow in 1–2 h. The Coriolis effect was not included, thus making this a truly two-dimensional simulation. A very similar initialization was used by Parsons (1992) to simulate this frontal system in two dimensions using a model with warm rain microphysics.

3. Results of the control simulation and comparison to observations

Using the experimental design described above, the numerical model was integrated for 8 h. Within 2 h, the modeled storm became quasi-steady. In most of the figures, the modeled storm is illustrated at t = 8 h. The evolutionary features of this storm will be presented in subsequent sections. At t = 8 h, the model-derived radar reflectivity pattern and airflow field (Fig. 2a) showed a strong updraft of up to 19 m s\(^{-1}\), producing a core of precipitation about 6 km wide. Most of the air continued to flow to the rear of the storm at midlevels, producing an extensive stratiform region composed mostly of snow and graupel. This airflow and reflectivity field agree well with the airflow and reflectivity patterns observed by Carbone (1982, see his Figs. 6 and 7). Additionally, both the magnitude and height (1.875–2.125 km) of the peak updrafts compare reasonably well to the average maximum vertical velocity of 17 m s\(^{-1}\) updrafts observed and the altitude of the observed maximum vertical velocity (2.1 km).

The structure of the storm (Fig. 2b) showed cloud water in the updraft core and rain mostly in the updraft core, but rain was also formed in the stratiform region from melting snow and graupel. The highest concentrations of snow occurred at the top of the convection and were then advected to the rear of the storm, creating the stratiform region. Likewise, graupel was formed at the top of the convection and was advected rearward, but the graupel also fell, producing significant precipitation on the surface 5 km rearward of the updraft core. Ice was widespread at upper levels.

The precipitation fell as rain in the convective region (Fig. 2c), reaching rates of up to 50 mm h\(^{-1}\). This model-derived precipitation rate was lower than the precipitation rate of over 100 mm h\(^{-1}\) for the convective region reported by Carbone (1982). In the stratiform region the precipitation was composed of rain and graupel and had rates of 1–18 mm h\(^{-1}\) over a 40-km distance. Carbone reported a precipitation rate of 2.5 mm h\(^{-1}\) for a 40-min stratiform region. Carbone found no evidence of graupel or hail reaching the surface in this storm. The model estimated that the amount of precipitation that fell from the storm was 8.10 mm (assuming the storm movement was 21.6 m s\(^{-1}\)). This is well within the range of measurements (6.7–14.5 mm) reported by Carbone.

In the updraft core, cloud water was produced through condensation and reached concentrations of up to 2.5 g kg\(^{-1}\) (Fig. 3a). The cloud water was transformed, through coalescence and accretion, to form rain, graupel,
and, to a lesser extent, snow. Rain-accreting cloud water (Fig. 3b), which occurred in the middle to lower regions of the cloud water, and graupel-accreting cloud water, which occurred in the mid- to upper regions of the cloud water, were the major sinks of cloud water. The rain field was the primary precipitation field in the convective region, reaching concentrations of 1.7 g kg⁻¹ (Fig. 4a).

The source of rain in the convective region was coalescence and accretion of cloud water, and the primary sink of rain was rain that had collected ice and became graupel-like in form (Fig. 4b). Rain was also formed in the stratiform region where it reached concentrations of 0.2

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Fig. 2. (a) Radar reflectivity (dBZ) derived from the model at t = 8 h and storm-relative airflow field velocity vectors. The maximum vector length is 25.5 m s⁻¹. (b) Hydrometeor fields after 8 h of model integration. Lightly stippled area is ice in excess of 1 µg g⁻¹. Darkly stippled area is snow in excess of 0.1 g kg⁻¹. Area with vertical lines is rain in excess of 0.1 g kg⁻¹. Dashed contour line is graupel equal to 0.1 g kg⁻¹. Heavy contour line is cloud water equal to 0.1 g kg⁻¹. (c) Surface precipitation rate at t = 8 h.

Fig. 3. (a) Cloud water mixing ratio (g kg⁻¹) at t = 8 h. (b) Sinks of cloud water at t = 8 h. Solid contours are coalescence plus rain-accreting cloud water. Dashed contours are graupel-accreting cloud water. Contours are 1, 5, and 10 (10⁻³ g kg⁻¹ s⁻¹).

Fig. 4. (a) Rain mixing ratio (g kg⁻¹) at t = 8 h. (b) Sources and sinks of rain at t = 8 h. Solid contours are coalescence plus rain-accreting cloud water. Dotted contours are rain collecting ice to form graupel. Contours are 1, 5, and 10 (10⁻³ g kg⁻¹ s⁻¹). Long-dashed contours are melting from snow and graupel with contours of 0.1, 0.5, and 1.0 (10⁻³ g kg⁻¹ s⁻¹).
g kg\(^{-1}\) from melting graupel and snow. Snow grew at the top of the convection to concentrations of 2.3 g kg\(^{-1}\) (Fig. 5a). The snow was advected rearward to the stratiform region where concentrations reached 0.5 g kg\(^{-1}\). The snow was formed from ice aggregation and grew through riming of cloud water, collecting ice crystals, and vapor deposition (Fig. 5b). The major sinks of snow were the formation of graupel from rime snow crystals, and sublimation and melting (Fig. 5c). Graupel was created in the convective region where concentrations reached 3.9 g kg\(^{-1}\) and was advected to the stratiform region where concentrations reached 1.5 g kg\(^{-1}\) (Fig. 6a). The primary processes that created graupel were snow-collecting cloud water and rain-collecting ice (Fig. 6b). Graupel grew in the convective region mostly through accretion processes but also through vapor deposition. In the stratiform region graupel was depleted through sublimation and melting (Fig. 6c). Ice was formed in primarily two layers: near the height of 6 km and above 9 km. Ice formation above 9 km occurred at the coldest point of the initial temperature sounding (Fig. 1), which easily became saturated with respect to ice and produced the high ice anvil seen in Fig. 2b.

Carbone (1982) estimated a precipitation efficiency as the average amount of precipitation observed at the eight rain gauge sites divided by the amount of condensate produced during an effective residence time. In his calculation of condensate production rate, he assumed that all the saturated water vapor that was lifted condensed and that there were no losses such as evaporation or sublimation in downdrafts. Using this calculation, Carbone determined that the precipitation efficiency was 77%. He further estimated that if there were losses of condensate due to entrainment and moist adiabatic descent, the precipitation efficiency was 70%.

The model-derived precipitation efficiency was estimated in two methods designated PE1 and PE2. PE1, defined as the amount of precipitation at the surface divided by the amount of water vapor transformed to the liquid or solid phase, was found to be about 53% over the 8-h integration. PE2, defined as the amount of precipitation at the surface divided by the amount of water vapor flowing into the model domain in the lowest 3 km at the right boundary, was found to be 54% over the 8-h integration. Because the definition of PE2 is similar to the method that Carbone used to calculate the precipitation efficiency, PE2 will be the precipitation efficiency reported in the following sections. As with the maximum precipitation rates being different between model results and observations, the precipitation efficiencies showed a similar difference. A number of possible explanations exist for the differences in the magnitude of the precipitation efficiencies in the observed and modeled systems, including errors in the observed rainfall rates or water vapor profile, violation of the sublimation/evaporation assumptions used by Carbone, three-dimensional effects not accounted for in the model, errors in the model formulation of the microphysical processes, and shortcomings in the initialization. However, the model estimates are more typical of precipitation efficiencies in convective systems (Fankhauser 1988) and within the range of precipitation efficiencies (0.53–0.67) estimated in the Rutledge (1989) study.

4. Effects of latent heating and cooling

To assess the effect of ice on the dynamics and thermodynamics of a winter cold front, simulations were also conducted with no ice and with no graupel. For
the simulation without ice, the peak updrafts were located between 1.875 and 2.375 km and reached speeds of 17 m s\(^{-1}\) (Fig. 7a). Again, the magnitude and location of the maximum vertical velocity agreed well with observations. The radar reflectivity derived from the model results showed a strong return in the core of the updraft; however, no stratiform region was produced. The structure of the storm for the simulation is shown in Fig. 7b. For the simulation with no ice, the cloud water occurred in a narrow core at lower levels where it reached concentrations of 2.1 g kg\(^{-1}\). The cloud water also existed at higher altitudes where it was widespread and had concentrations of 0.1–0.5 g kg\(^{-1}\). The rain was restricted to the updraft region where it reached concentrations of up to 5 g kg\(^{-1}\). At \(t = 8\) h, the precipitation rate (Fig. 7c) for the simulation with no ice was over 100 mm h\(^{-1}\) in the convective region and 0 mm h\(^{-1}\) in the stratiform region. The precipitation rate in the convective region replicates the observed precipitation rate in the convective region, but the nearly complete lack of stratiform precipitation indicates that the ice phase needs to be included in the
model to replicate the stratiform region in this system, as it has been noted in previous studies of observations of stratiform rainfall in mesoscale convective systems (Zipser 1969; Houze 1977) and replicated in numerical simulations (Fovell and Ogura 1988). In the observations of this front, the cold air mass behind the heavy rainfall deepened very slowly with time. In cases where the cold air deepens more rapidly, the large-scale ascent of air over the frontal surface may produce a stratiform region even in the absence of ice.

For the simulation conducted without graupel, an extensive stratiform region was formed, as observed in the radar reflectivity pattern and the location of the hydrometeor fields (Figs. 8a and 8b). The peak updrafts in this simulation were 19 m s$^{-1}$ and occurred from 1.875 to 2.375 km. The main difference between the no-graupel simulation and the control simulation is that in the no-graupel simulation the snow field was much greater in concentration (up to 3.8 g kg$^{-1}$ at the top of the updraft core and 0.1–2.0 g kg$^{-1}$ in the stratiform region) and was more extensive. The surface precipitation rate for the no-graupel simulation (Fig. 8c) was 55 mm h$^{-1}$ in the convective region and about 12 mm h$^{-1}$ in the stratiform region. The model-derived total precipitation rate for this simulation was less than the precipitation rates observed. The precipitation efficiencies of 60% for the simulation with no ice and 54% for the simulation with no graupel were again somewhat less than that calculated for the observed storm by Carbone (1982).

These simulations seem to capture many of the basic features of the observed system. However, it is a difficult task to replicate all observed characteristics of a storm, including surface rainfall rates, radar reflectivity, and derived airflows, with a parameterized model that incorporates a simple initialization. It becomes more difficult as the number of predefined parameters in the model increases. For example, to describe the size distribution of snow (see Rutledge and Hobbs 1983), the model parameterization assumes a constant $N_{a0}$ of 0.2 cm$^{-4}$ and allows $\lambda_s$, the slope of the distribution line, to vary (for the stratiform region in the model simulation $\lambda_s$ varied between 20 and 30 cm$^{-1}$). Measurements of hydrometeors from a cold-frontal rainband in a mature frontal system observed in the central valley of California on 13 March 1984 indicate that both $N_{a0}$ and $\lambda_s$ vary from 0.04 to 0.57 cm$^{-4}$ and from 12 to 50 cm$^{-1}$, respectively (Gordon and Marwitz 1986). While the model parameters are within the range of values observed, it may not be possible to predict accurately the observed variations in the snow distribution. However, the model captures the dynamics of the storm fairly well. The basic airflow, the magnitude of the vertical motion field, and the radar reflectivities compare well between model results and observations. However, comparisons of the observed surface precipitation rates to the results of the three simulations indicate that it is difficult to replicate quantitatively the surface rainfall rates within both the heavy band of rainfall and the zone of trailing stratiform rainfall. To improve this model result, some effort needs to be made in improving the ice microphysical parameterization. In particular, the division of particles into two precipitating ice types such as snow or graupel is unable to represent the spectra of rimed particles found in nature.

Despite these slight discrepancies between model results and observations, the model results can provide
some insight into how the microphysical processes influence the dynamics of the cold front. The primary interaction between microphysical processes and frontal dynamics occurs through the latent heating and cooling as hydrometeors condense, evaporate, freeze, melt, or sublimate. Throughout the model integration of all three simulations, the condensational heatings that occurred in the core of the updraft were all about the same, reaching values of over 180 K h\(^{-1}\). Since the lapse rate ahead of the rainband was nearly neutral to moist ascent, this large diabatic heating was offset by adiabatic cooling during vertical ascent. However, the two simulations with ice microphysics had additional heating in the updraft core due to freezing processes and very different distributions of diabatic cooling.

The cold pool, defined as the region behind (to the left of) the updraft core, is important for developing and maintaining circulations within the storm (e.g., Rotunno et al. 1988). For the three simulations discussed in this study, the cold pool cooled by 1–6 degrees (Fig. 9), with the greatest cooling taking place during the control simulation. The magnitude of the cooling rate from diabatic processes (−7.2 K h\(^{-1}\)) in the cold pool was nearly the same for the three simulations, but the location, extent of coverage, and evolution of the cooling differed (Figs. 10–12). In the control simulation (Fig. 10), latent cooling from sublimation occurred from the freezing level (at 0.9 km at \(t = 8\) h) to about 2.5 km in height and extended 20 km rearward from the updraft core. Most of the latent cooling occurred at earlier times in the model run. At \(t = 2\) h, the latent cooling had a strong band of cooling at the 1.5-km height. Above 1.6 km (the freezing level) the cooling was due mostly to sublimation and some evaporation, but below the freezing level most of the cooling was from evaporation of melted particles at this early time in the model integration. This band of cooling decreased in magnitude as the model integrated forward, and by \(t = 8\) h, most of the cooling was centered at \(z = 1.5\) km and was less than 10 km from the updraft core.

In the simulation without ice (Fig. 11), the cooling was limited to a 6-km area just to the left of the updraft. The evolution of the latent cooling for this simulation was smaller at \(t = 2\) h and grew in strength and somewhat in area with time. The cooling for the simulation with no graupel (Fig. 12) occurred mostly between 2

Fig. 10. Latent heating and cooling associated with the diabatic processes in the control simulation. Contours are −7, −5, −2, −1, 5, 10, 20, and 50 K h\(^{-1}\): (a) at 2, (b) 4, (c) 6, and (d) 8 h.

Fig. 9. Potential temperature deviation from the initial state averaged in an area of the cold pool that was 1 km deep and 10 km wide nearest the updraft core. At \(t = 0\), the potential temperature deviation of −6°C was imposed in the cold pool to simulate a density current. Curve a is the control simulation, curve b the simulation without ice, and curve c the simulation with no graupel.
In the two simulations where ice was included, the magnitude of both the updraft and the left to right flow in the cold air increased. The updrafts increased in magnitude by about 2–3 m s$^{-1}$ (Fig. 13) in the simulations with ice, which is consistent with the greater intensity of the cold pool increasing the forced ascent and higher water loading in the no-ice simulation near the updraft. The left to right flow in the cold air was considerably influenced by this latent cooling. For the simulations with ice and with no graupel, the magnitude of the left to right flow in the cold air peaked at

![Fig. 11. As in Fig. 10 but for the simulation without ice.](image)

and 3 km in height (above the freezing level) within the stratiform region. Unlike the control simulation, the latent cooling in the stratiform region became more widespread with time but did not change in magnitude. Another characteristic shown in the no-graupel simulation is the horizontal stratification of latent cooling and heating in the mid- and upper levels of the stratiform region. Small updrafts and downdrafts were associated with these layers, possibly due to buoyant motions within destabilized layers. This layering effect also occurred in the control simulation but it was not as pronounced.

![Fig. 12. As in Fig. 10 but for the simulation with no graupel.](image)
over 8 m s$^{-1}$ (Fig. 14), whereas for the simulation with no ice the magnitude of this airflow reached only 6 m s$^{-1}$.

5. Description and discussion of the sensitivity tests

Three sensitivity tests were performed to evaluate the response of the dynamics of the storm to different initialization scenarios. First, the model was run in a no-shear environment to estimate the importance of the shear. Second, the depth of the initial cold air was altered to test whether the height of the cold air in density current circulations is determined by the height of the melting layer. Finally, the effect of decreased humidity in the initial cold pool was evaluated.

a. Effect of a no-shear environment

The vertical shear of the horizontal wind in the control simulation was 20 m s$^{-1}$ over 3 km and played an important role in maintaining the narrow cold-frontal rainband. To test the importance of this shear as was proposed in previous studies and to study how the previously observed diabatic cooling depended upon the shear, a model simulation with no initial environmental shear was performed. The primary difference between the simulation with no shear and the control simulation is that the simulation with no initial shear was generally weaker (Fig. 15), with smaller updrafts (10.5 m s$^{-1}$) and less precipitation in the convective region (cf. Figs. 2 and 15). The no-shear simulation also had higher reflectivities in the stratiform region and cooled less (−6°C initially to −6.7°C at 8 h). Thus, the freezing level did not descend with time and after 8 h of integration was much higher than in the control simulation. This result suggests that the local diabatic cooling may be less effective in the no-shear simulation, although the very broad anvil and the limited domain employed in this study (100 km) do not preclude the possibility of cooling taking place near the rear edge of the stratiform rain region.

The microphysical processes were different in the simulation without vertical shear as expected given the decreased intensity of the updrafts. In general, the role of stratiform ice processes increased, such as deposition onto ice particles, conversion of ice to snow, and accretion of ice by snow. The melting also occurred over a larger area in the no-shear simulation. However, the cooling from the melting and evaporation in the no-shear simulation was equal in magnitude to the cooling from sublimation and evaporation processes in the control simulation. The interpretation of this finding is straightforward in view of past studies (Parsons 1992) that suggest stronger deeper ascent takes place in the presence of vertical shear through a dynamic interaction between the vertical shear and the cold pool. The shallower, sloped ascent in the no-shear simulation produces more ice particles and a broader stratiform rain area.

b. Factors that control the cold pool height

Carbone (1982) speculated that the height of the cold air mass was determined by the height of the melting layer. However, Locatelli et al. (1995) did not find a correspondence between melting layer height and height of the cold air mass in their study of narrow coldfrontal rainbands observed at Cape Hatteras, North Carolina, during the winter. In a theoretical study, Xu et al. (1996) proposed a dynamic control on the depth of the cold air given an infinite reservoir of cold air so that the height of the cold air mass corresponds to the height of the environmental shear. To investigate further the factors influencing the height of the cold air mass in cases where these rainbands are present, sensitivity tests were performed where the initial height of the cold pool was altered.

The variations in the height of the freezing level and the $\theta' = -3^\circ$C contour are given in Fig. 16, where $\theta'$ is defined as the deviation of potential temperature from its initial state. The $\theta' = -3^\circ$C contour was selected to indicate deepening of the cold air mass. In this illustration, the height of the freezing level is representative of the average freezing level height for the region from the left model domain edge to the updraft, and the height of the $\theta' = -3^\circ$C is representative of a 10-km subdomain centered 10 km to the left of the updraft. In the control simulation, the freezing level descends with time, while the depth of the cold air mass increases with time (Fig. 16). The result again reinforces the theme of a gradual intensification of the cold air mass that takes place through ice phase processes. A smaller, but still noticeable, amplification of the cold air mass takes place in the simulation without graupel (Fig. 16). In the control and no-graupel simulations, the depth of the cold air mass does not correspond to the height of the freezing level as proposed by Carbone (1982) and
Rutledge (1989), as both variables change as a function of time. The smaller changes in the simulation without graupel suggest that this result is to some degree dependent on the microphysical parameterization. Without an ice phase, the changes are far less pronounced.

To further investigate this gradual deepening of the cold pool with time and the relationship of the freezing level to the height of the cold pool, we conducted a simulation where the initial depth of the cold air mass was reduced from 5 to 3 km. This simulation used the same full microphysical scheme as the control simulation. The results from this simulation showed that the cold pool height once again increased as the freezing-level height decreased with time (Figs. 17a and 17b). In the simulation where the initial height of the cold pool was set to 3 km, the height of the cold air mass after 8 h of integration was level at 3 km, and the location of the melting layer in the cold air dropped to about 1.2 km (Fig. 17b). The control simulation is shown in Figs. 17c and 17d. The difference in the depth of the cold air mass is relatively small between the two

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**Fig. 14.** Horizontal velocity at $t = 8$ h for (a) the control simulation, (b) the simulation without ice, and (c) the simulation with no graupel. Units are m s$^{-1}$.

**Fig. 15.** As in Fig. 2 but for the simulation with no shear. The maximum vector length is 28.7 m s$^{-1}$. 
cases, especially in regard to the larger differences between the initial states. For another simulation with the control microphysics where the height of the initial cold pool was set to 2 km, the height of the cold air mass after 8 h of integration ranged from 1.75 km near the convective region to 2.75 km near the domain edge (60 km away), and the freezing level was at 1.6 km. This simulation also showed the characteristic deepening of the cold pool with time.

This comparison suggests that a possible self-intensifying mechanism for the circulations associated with these rainbands exists. It is interesting to note that in the cases when the cold pool was initialized at 3 and 5 km the cold pool depth seemed to reach a steady state. This finding is consistent with the idea that the depth of the cold pool tends to reach a quasi-steady state that depends on the dynamics of the ambient flow (Xu et al. 1996). Xu et al. (1996) proposed that given an infinite reservoir of cold air the environmental shear determines the height of the cold air mass. For our simulations, the initial wind profile had a strong shear layer from the surface to 3.1 km and a weak shear layer above 3.1 km (Fig. 1). Our results tend to support the theory of Xu et al. (1996). However, Xu et al. did not address the no-shear case. As stated earlier, for our simulation that had no shear initially, the intensification of the cold pool was weaker. It is clear that further research is needed in order to investigate how their idealized study applies to these atmospheric flows.

c. Effect of the humidity in the cold pool

Because it appears that the diabatic cooling within the cold pool affects the dynamics of the narrow cold-
6. Discussion and comparisons to previous work

In the results section, comparison of the model results with observations were presented. Overall, the comparison was favorable. It is also useful to evaluate how these model results compare with other previous studies. Rutledge (1989) used a microphysical cloud model with a fixed airflow to examine the precipitation mechanisms of the winter storm examined in this study. There are four major differences between the model used by Rutledge (1989) and the model that we used. In the Rutledge study the model was three-dimensional, used a fixed airflow, made the hydrostatic assumption to derive a thermal field, and used an upstream advection and precipitation flux calculation. The model we used was a two-dimensional, dynamic, nonhydrostatic model that used centered-space, leapfrog advection and precipitation flux calculations. The microphysical pa-

frontal rainband, the dependency of this cooling on the thermodynamic characteristics of the cold air mass was investigated. Recall that the control simulation had a high relative humidity in the cold pool (91%). A simulation was performed that had a cold air mass with lower humidity (12% of the environmental relative humidity). This reduction in relative humidity produced a cold pool relative humidity of 16.4%. After 8 h of simulation, the model results showed a storm with many characteristics similar to the control simulation. The peak updrafts were 22 m s\(^{-1}\). The precipitation rate had a similar structure, although the peak precipitation rate in the low humidity simulation was larger (57 mm h\(^{-1}\)). The precipitation efficiency (54%) for the low-humidity simulation was about the same as the control simulation. The primary difference between these two simulations occurred in their evolution. The cooling in the cold pool for the low-humidity simulation was rapid in the first 1.5 h of the simulation, whereas the cooling in the control simulation was more linear (Fig. 18). This suggests that rapid cooling occurs until the cold pool is near saturation (85% RH or greater); then the cooling proceeds at a more gradual rate. The stronger updrafts in the low-humidity simulation tended to produce more precipitation in the convective region and less precipitation in the stratiform region, where increased sublimation occurred.

Fig. 17. Deviation of potential temperature from its initial state for the simulation where the initial cold air mass depth was 3 km for (a) \(t = 2\) h and (b) \(t = 8\) h, and for the simulation where the initial cold air mass depth was 5 km for (c) \(t = 2\) h and (d) \(t = 8\) h. The solid line is the freezing level.

Fig. 18. Potential temperature deviation from the initial state averaged in an area of the cold pool that was 1 km deep and 10 km wide nearest the updraft core. Curve a is the control simulation, and curve b is the simulation with initially low humidity in the cold air mass.
rameterization was the same except for minor changes as noted in the model description.

The microphysical results from Rutledge’s (1989) work and from this study are fairly similar. However, some minor differences exist. For example, the condensation process in the control simulation produced more cloud water than did Rutledge’s simulation. This difference in turn affected the magnitude of the accretion of cloud water by graupel and rain. The other main difference is that the control simulation performed for this study had much greater amounts of deposition of water vapor onto graupel and sublimation of graupel particles than Rutledge (1989) reported.

An exception between the Rutledge (1989) study and our work is that some graupel reached the ground in our control simulation. It is possible that this difference between graupel fields could be explained by the two-dimensional limitation of our model. Rutledge found that graupel was advected along the front, while our simulation did not include the along-front winds because of its two-dimensionality. Future work considering ice microphysics in a three-dimensional cloud model should certainly be done for these type of severe cold-frontal rainbands.

The differences between the graupel fields in Rutledge’s study and our work may certainly be due to differences in the airflow and thermal fields in the two models. A pronounced difference is found in the thermal fields, as Rutledge (1989) predicted that the highest temperatures were located at and slightly behind the observed surface front with a warming of several degrees (see Fig. 12 in Rutledge 1989), while we observed a sharp cooling at the leading edge of the front (Fig. 19). The question that naturally arises is which thermal field is more realistic. Rutledge proposes that the predicted warming is consistent with 1) subsidence behind the leading edge of the front and 2) the delay between the temperature drop and the windshift/pressure rise observed near the leading edge of these intense frontal rainbands (e.g., Hobbs and Persson 1982). However, we believe that the warming predicted to occur near the leading edge of the front in the Rutledge (1989) study is unrealistic because the thermodynamic energy equation used in Rutledge (1989) assumed a hydrostatic state (and furthermore assumed $\partial p'/\partial t = u(\partial p'/\partial x) = 0$), which was found not to be true (Carbone 1982; Parsons et al. 1987) because of the strong vertical motions being generated in vertical scales of 1–2 km and the large pressure jumps (3.5–5.5 mb) observed (Carbone 1982). Since the equation used by Rutledge (1989) is unable to predict the non-hydrostatic pressure forces that drive the vertical motions in this region, the model will adjust the thermal field accordingly; hence, we believe that the derived warming in the Rutledge (1989) study may be an artifact of the assumptions used in that model. Furthermore, while surface measurements (Hobbs and Persson 1982) do show a time lag between the pressure jump/ windshift location and the temperature drop consistent with a nonhydrostatic flow, pronounced warming at the front has not been observed. Instead, vertical cross sections (Bond and Fleagle 1985) show pronounced local cooling just behind the surface front.

A final reason to discount the thermal fields near the leading edge of the front in the previous kinematic study is the vertical motions in the region may be suspect because the airflow field was derived from Doppler analysis. Besides the many assumptions that go into the Doppler analysis, it was necessary to fill in the horizontal and vertical wind data in the lower levels where raw data were missing. To do this, the lower boundary condition was set to $w = 0$ at the surface, and the continuity equation was used to fill in the missing data near the surface and integrated upward. As a result of this Doppler analysis, the flow field that was used by Rutledge (1989) showed downdrafts in excess of 10 m s$^{-1}$ behind updrafts of 25 m s$^{-1}$. A downdraft of this magnitude may overestimate the postfrontal subsidence that Carbone (1982) noted to be commonly about 5 m s$^{-1}$.

We suspect that the strong downdraft in Rutledge’s study may be a spurious result of his necessary filling of the radar data. Unfortunately, in a fixed airflow model there is not the opportunity for the model dynamics to overcome these imbalances.

What we believe to be anomalously warmer temperatures in the Rutledge (1989) simulation did allow for the complete melting of graupel prior to it reaching the surface, which is consistent with the observations. As noted earlier, the control simulation showed that graupel did reach the ground. We attribute this difference to be partly due to the different heights of the melting level in the two cases and the differences in the airflow. In this study the freezing level was as low as 1.1 km late in the simulation (Figs. 16 and 17), while in the Rutledge (1989) work the freezing level was as high as 2.75 km (see his Fig. 12) with the graupel melting commencing as high as 2.5–2.6 km (see his Fig. 18). Given the descent of the freezing level in the control simulation, one would expect that graupel would reach
the surface at later times during the simulation. However, we do concede that we may be underpredicting the melting of graupel since limited amounts of graupel do reach the surface early in our control simulation. For this reason, the slightly slower intensifying no-graupel simulation may provide a more conservative estimation of the amplification of the cold pool and the descent of the melting level. It is interesting to note that graupel also reached the ground in the Rutledge and Hobbs (1984) simulation of another intense frontal rainband, although to our knowledge no graupel was observed to reach the surface in that system (Hobbs and Persson 1982). This finding, and the results earlier in our study before the melting level dropped, lends us to suspect that this type of graupel parameterization may slightly overpredict graupel formation or underpredict their melting rates. A possible reason for this difference could be the lack of collisional breakup or a slower graupel formation in this type of formulation at temperatures slightly cooler than 0°C where graupel first begins to form.

Another potential reason as to why graupel reached the surface in the control simulation and did not in the results presented by Rutledge (1989) is the advection scheme employed in the two models. Rutledge's model used an upstream differencing advection scheme, whereas our model used a centered-space, leapfrog advection scheme. Larger gradients would occur in the centered-space method compared to the upstream method, thus producing a larger sedimentation flux. These large gradients could be suppressed if a fine vertical resolution were used. We performed a simulation with a 125-m vertical resolution (half the grid spacing of the control simulation) that produced similar results to the control simulation, including graupel reaching the surface. Thus, either an even higher vertical resolution should be used with the centered-space sedimentation flux calculation, or the centered-space advection scheme is not contributing largely to the graupel reaching the surface.

In the section describing the effect of latent heating, we found that sublimation of snow and graupel played an important role in the evolution of the narrow cold-frontal rainband that was modeled, as it increased the intensity and depth of the cold air mass. Additionally, it was found that the sublimation process contributed more to the latent cooling than did the melting process. Clough and Franks (1991) performed a detailed analysis of evaporative processes in frontal and stratiform precipitation. Their results also showed that sublimation of ice particles was an efficient thermodynamic process and that the mass of ice particles became the limiting factor for latent cooling processes. Huang and Emanuel (1991) first investigated the role of evaporation of rainfall upon cold-frontal circulations, showing that it significantly increases the rate of frontogenesis. Recently, Parker and Thorpe (1995) investigated the role of snow sublimation on frontogenesis. Although their study showed that this diabatic cooling had little effect on frontogenesis and on the large-scale flow in their semigeostrophic model, the cross-frontal flows in the vicinity of the sublimation were strongly modified. In particular, a mesoscale downdraft was produced below the synoptic frontal surface in a manner consistent with the data from the FRONTS 87 experiment. As seen in the results of our simulations, the previous work of Parker and Thorpe (1995), and in previous numerical simulations (Fovell and Ogura 1988) and in observations (Zipser 1969; Houze 1977), the inclusion of an ice phase is necessary to accurately represent the mesoscale circulations and precipitation structure in the vicinity of cold fronts.

The techniques of Parker and Thorpe (1995) and our study represent two distinctly different approaches to the investigation of the impact of ice processes and the associated diabatic cooling on cold fronts. The Parker and Thorpe (1995) study is idealized but in a sense it is more general, as it is aimed at the impact of ice on the scale collapse of a semigeostrophic cold front. Their study is also on a larger scale, and the diabatic cooling needs to be parameterized. Our study is aimed at smaller-scale processes through directly resolving convective-scale motion. The work is aimed at a specific type of intense frontal system as described initially in Browning and Harrold (1970), Carbone (1982), Hobbs and Persson (1982), Bond and Fleagle (1985), and in subsequent studies. The density current model is an idealization of these fronts; however, not all cold-frontal systems conform to this model nor is the idealization without fault (Smith and Reeder 1988; Parsons 1992).

Our finding that the diabatic cooling due to ice processes increases the depth and intensity of the cold air mass is relatively robust in that it appeared in both the control and no-graupel simulations and in those simulations where the depth of the cold pool was varied. It also occurred in simulations where the vertical resolution was doubled (not shown). These results support the hypothesis that the diabatic cooling due to the ice phase may intensify or self-sustain these intense frontal squall lines against the dissipative effects of surface fluxes, mixing, and rotation. A point in favor of this self-intensifying mechanism is that shallow and weak cold air masses are generally not observed in mature precipitating, maritime cold fronts observed in detail with aircraft and Doppler radar; rather, circulations similar but less intense than the Carbone (1982) case seem to be common [see review of observed fronts in Parsons (1992)]. The cold pool depths are also deeper than commonly observed, with more buoyant, warm season convective outflows, despite the fact that a stronger more vigorous downdraft can occur in these warm season events.

As stated earlier, previous studies (Carbone 1982; Rutledge 1989; Moncrieff 1989) argued that the top of the cold air mass should correspond to the melting level.
and further hypothesized that the melting inhibited mixing by increasing stability across the cold air interface. The mixing would most likely be due to Kelvin–Helmholtz waves traveling rearward from the leading edge of the cold air interface. The grid spacing used in the current simulations, while very fine scale compared to traditional studies of fronts, cannot resolve this wave process. As stated earlier, increasing the vertical resolution did not change the finding that the cold air mass deepened and intensified, although a distinct melting inversion was observed in this higher-resolution simulation. The caveats of not resolving waves and mixing being parameterized leave open the possibility that suppression of mixing across the melting level may play a role, although our results tend to disfavor mixing in favor of diabatic cooling within the cold air mass. The local thermal minimum behind the leading edge of the cold front as noted by Bond and Fleagle (1985) and Parsons et al. (1987) support this theory that diabatic processes influence the depth of the cold pool.

With the previously stated caveats in mind, our results do not support the interpretation of melting as playing a primary role in marking the top of the cold air mass and inhibiting mixing. Our results have led us to reexamine the evidence presented in the Carbone (1982) study for this argument. The strongest points in favor of the previous viewpoint are that the temperature difference between the pre- and postfrontal soundings reversed just above the melting level and that the layer just above the melting level was very stable (i.e., isothermal). Examination of the postfrontal sounding, however, clearly shows a second temperature inversion between 3.1 and 3.5 km (see Fig. 17 of Carbone 1982). The top of this inversion layer also corresponds to the region where there is strong backing of the horizontal wind with height (see Fig. 18b of Carbone 1982). The backing of the wind, through the thermal wind relationship, indicates that the level of cold advection extended to \( \sim 3.6 \) km. These two points seem to suggest that the top of the cold front may be nearer the 3.1–3.6-km level, rather than the 2.1-km level. In the control simulation of our study, pre- and postfrontal soundings showed similar features to those illustrated by Carbone (1982), except for the height of the freezing level (Fig. 20). Another piece of evidence for a deeper cold air mass than suggested by Carbone (1982) is that the “kinematic cold front,” the depth of the isosurface of zero relative velocity, was near 2.1 km. The relative flow within a density current suggests that the true cold air boundary should be above this kinematic estimate (Moncrieff 1989). It is worth noting that in applying an analytical model of a density current to this case Moncrieff (1989) used a cold air mass depth of 3.4 km with the observed storm structure and movement and found good agreement. Hence, a number of pieces of evidence suggest that the melting level in the observations of this system may have actually been located well within the cold air mass and have little to do with blocking mixing of warm sector air into the cold air mass.

If our interpretation is correct, then the diabatic processes may even cause an intensification of the front with time rather than just maintaining a steady-state structure. A number of theoretical (e.g., Moncrieff and So 1989) and modeling studies (Parsons, 1992) have treated NCFR circulations as steady state. While the evolution implied here may have important consequences for maintaining these circulations, the evolution is rather gentle, as the changes take place over 8 h. Thus, it seems quite likely that this evolution does not invalidate the steady-state studies in terms of investigating the dynamics associated with the strong updraft.

The intensification of cold pools with time in precipitation systems is not a new finding, and such evolution can occur in simulations of active convection using warm rain microphysics (e.g., Rotunno et al. 1988). However, in many cases the strengthening of the cold pool may imply a weakening of the system (updraft strength, rainfall rate, etc.), as the density current circulations begin to strengthen and negatively impact the generation of deep convection, as the vertical shear is too weak to dynamically balance the cold pool vorticity (e.g., Weisman et al. 1988). The importance of the cooling in strengthening the circulations in this case is that the density circulation is the primary precipitation-producing feature of the system due to the intense shallow updrafts, relatively deep cold pools, and strong vertical shear associated with frictional turning of the low-
level jet in the presence of near-saturated conditions. In this case the prefrontal flow may have been channeled by orography, resulting in enhanced vertical shear. It is also worth noting that the mechanisms for the intensification, particularly sublimation within the cold pool, will generally not play a major role in most warm season convective events since the freezing level is so high above the height of the cold pool.

7. Conclusions

The microphysics and dynamics of a narrow cold-frontal rainband and its associated stratiform region were simulated using a two-dimensional prognostic cloud model that employs a bulk water microphysics parameterization. The general structure of the storm replicated observations fairly well, but there were some differences in precipitation efficiencies and precipitation rates between model results and observations, and some graupel reached the surface only in the simulations. A simulation without ice microphysics showed that a trailing stratiform region did not develop and, therefore, did not replicate the observations of the stratiform region well but did reproduce the convective region well.

The initial conditions of the model were altered to examine the effect of a no-shear environment, the height of the initial cold pool, and the humidity of the cold pool. In general, the storm that developed from the no-shear environment was weaker than the control simulation. By altering the height of the cold pool, we learned that the height of the cold air mass was not determined solely by the freezing level for the case studied. The height of the cold air mass is probably due to latent heating and cooling processes in the stratiform region behind the deep convection coupled with the environmental shear. By reducing the humidity in the cold pool, cooling in the cold pool occurred much more rapidly but eventually leveled off after approaching water saturation.

All these simulations seem to indicate that the addition of the ice phase induces cooling within the cold air mass that increases the baroclinicity across the front. A larger-scale investigation by Parker and Thorpe (1995) showed that the inclusion of cooling due to an ice phase was necessary to obtain realistic mesoscale downdrafts. The cooling in our study is associated with sublimation and melting and also induces a nonsteady aspect to these fronts not readily evident in simulations with only warm rain microphysics. This study also suggests that the ice phase may play a role in maintaining these types of fronts in the atmosphere as hypothesized from observations. In these simulations, the mechanism for the ice phase maintaining these fronts is not to decrease the mixing across the frontal air mass by the stabilization induced by melting, as hypothesized in earlier studies, but rather by the pronounced cooling itself. However, this result may still be considered preliminary, especially given the two-dimensionality and highly parameterized microphysics of the model.

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