Modeling convective-stratiform precipitation processes on a Mei-Yu front with the Weather Research and Forecasting model: Comparison with observations and sensitivity to cloud microphysics parameterizations

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[1] Deep convective-scale simulations of the linear mesoscale convective systems (MCSs) formed on a Mei-Yu front over the Huai River basin in China on 7–8 July 2007 were conducted using the Advanced Research Weather Research and Forecasting model to investigate impacts of cloud microphysics parameterizations on simulated convective-stratiform precipitation processes. Eight simulations were performed with identical configurations, except for differences in the cloud microphysics parameterizations. Measurements from rain gauges, ground-based weather radars, and the Tropical Rainfall Measuring Mission satellite Precipitation Radar were used to quantitatively evaluate the model results. While all of the simulations largely capture the observed large-scale characteristics of the precipitation event, notable differences among the simulations are found in the morphology and evolution of the MCSs at mesoscale and cloud scale. Significant influences on the coupling between dynamical and microphysical processes at the resolved deep convective scale by the various microphysical parameterizations are evident. On the one hand, the different microphysical schemes produce not only substantial differences in intensity of convective precipitation but also distinguishable vertical distributions of latent heating and condensate loading in the deep convective regions, which in turn results in significant differences in the vertical distributions of vertical air velocity and in the heights and strength of detrainment from deep convective regions. Consequently, detrainment of hydrometeors and positively buoyant air from the deep convective regions to the stratiform regions is significantly different, which impacts the formation and growth of ice-phase hydrometeors at the upper levels and thus surface rainfall rates in the stratiform regions. On the other hand, prediction of rain size distribution significantly impacts the simulated rain evaporation rates and mass-weighted rain fall speeds, and hence rain flux. Improper determination of the intercept parameter of rain size distribution can result in unrealistic features in the morphology of the storm and can have substantial impacts on precipitation distribution and evolution.


1. Introduction

[2] Because of the need for a unified formulation of the entire spectrum of the interactions among various dynamical and physical processes across the range of relevant spatial and temporal scales for more accurate weather and climate prediction [Arakawa, 2004], and with the rapid growth of computational capacity, the grid resolutions used to run atmospheric numerical models have been increased over the past four decades. More and more simulations and predictions are performed at the “cloud-resolving” or ‘explicit-deep convection’ scales, i.e., with a grid spacing of a few kilometers in the horizontal and a few hundred meters in the vertical. Currently, continental-scale numerical weather prediction (NWP) can be performed at cloud-resolving scales (e.g., the Weather Research and Forecasting (WRF) model [see Skamarock et al., 2007]). For climate modeling, cloud-resolving models (CRMs) have been implemented at each grid column of the general circulation models to replace

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Figure 1. (top) Distribution of the rain gauge (dots), radar stations (circles), and Wangjiaba reservoir (cross) within the analysis domain, as well as coastline (solid line). (bottom) Huai River and Yangtze river (thick solid lines) and boundaries of provinces (thin solid lines); shadings represent the terrain height.

most of the traditional parameterizations [e.g., Khairoutdinov and Randall, 2001; Grabowski and Smolarkiewicz, 2002; Tao et al., 2009], and global versions of CRMs have emerged [Tomita et al., 2005; Miura et al., 2007]. Such high resolutions are able to (at least partially) explicitly represent deep convective-scale and mesoscale dynamical processes and thus avoid the use of parameterization of moist convection, which remains probably the largest uncertainty in current NWP and climate models. However, cloud microphysical processes, as well as other important processes such as turbulence and radiative transfer, still need to be parameterized, and parameterization of these processes introduces uncertainties in models at the cloud-resolving scales [Tao and Moncrieff, 2009].

Despite the higher grid resolutions and more sophisticated physical parameterizations in atmospheric numerical modeling, significant deficiencies in model precipitation and the fine-scale structures of mesoscale convective systems (MCSs) are often present. For example, CRMs have not been very successful at replicating commonly observed squall line features, the transition zone and trailing stratiform region in particular [Fovell and Ogura, 1988; McCumber et al., 1991; Sui et al., 1998; Lang et al., 2003; Tao et al., 1993, 1995]. Cloud-resolving simulations of precipitation events have suggested a higher convective-stratiform rain ratio or an overestimate (underestimate) in the frequency of occurrence of large (small) radar reflectivity near the surface [Zhou et al., 2007; McFarquhar et al., 2006].

[4] The Mei-Yu is a climatological phenomenon affecting a large area in eastern Asia and producing a large amount of rainfall in the region [Ding, 1994]. The Mei-Yu season is characterized by a quasi-stationary front (the Mei-Yu front, or Baiu front) on surface weather maps which separates the Pacific subtropical high to the south from the migratory high over central China. At any given time, the Mei-Yu front is associated with the presence of an elongated cloud band, which is usually a mixture of deep convective clouds and stratiform clouds, extending from the interior of southern China to southern Japan on satellite images.

[5] The WRF model is currently used as a modeling tool in many countries around the world; the model is used not only for operational forecasting but also for research. To our knowledge, however, there have been few studies focusing on an evaluation of the WRF model’s performance in terms of cloud-resolving-scale modeling of the convective-stratiform processes in MCSs occurring on the Mei-Yu front in mainland China. This type of evaluation is one objective of our present study. The other, more important, objective is to examine the impacts of cloud microphysics schemes on the simulated convective-stratiform precipitation processes. This examination is motivated by previous studies which have suggested that cloud microphysics have a significant influence on precipitation structure in high-resolution simulations [McFarquhar et al., 2006; Lin and Colle, 2009; Jankov et al., 2009; Morrison et al., 2009].

[6] During 7–8 July 2007, heavy rainfall events occurred on the Mei-Yu front over the Huai River valley. This study focuses on convective-stratiform precipitation processes that occurred during a 13 hr period from 1700 UTC 7 to 0600 UTC 8 July, when the MCSs on the Mei-Yu front evolved from formation to dissipation and exhibited a clear structure of leading convective lines and trailing stratiform precipitation regions at their mature stage (2300 UTC 7–0200 UTC 8 July). Section 2 introduces the observational data sets, section 3 describes the case, and section 4 describes the design of the standard and experimental simulations and the analysis method. In section 5, precipitation and radar reflectivity from the standard simulations are compared with the observational data sets. The analyses of stratiform-convective processes in the standard simulations are presented in section 6, and results from the experimental simulations are discussed in section 7, with a concluding summary in the final section.

2. Data Sources

[7] The thermodynamic features of the Mei-Yu front and the large-scale flow in the vicinity of the Mei-Yu front are investigated using National Centers for Environmental Prediction (NCEP) FNL (Final) Operational Global Analysis data, which are updated on 1.0 × 1.0 degree grids continuously every 6 hours. The analyses are available on the surface and 26 pressure levels from 1000 mb to 10 mb.
Within the analysis region, there are 3195 surface meteorological stations providing rain gauge records and 10 operational Doppler radars deployed at the meteorological stations (Figure 1). These radars, operating at 10 cm wavelength, performed volume scans with nine elevation angles at 6 min intervals. The reflectivity data at each radar station were quality controlled to remove noise and interpolated from polar to Cartesian coordinates. The data were then combined to cover the full analysis region. For locations within the analysis region that were covered by multiple radars, an exponential weighting function of distance between the radar and the target place was used to take into account the measurements from the multiple radars. The horizontal grid interval was 1 km. The vertical grid interval varied with height, being 0.5 km at altitudes of 0.5–6 km and 1 km at 6–18 km. Horizontal distributions of the hourly accumulated precipitation rates derived from the rain gauge records were carefully compared with those of radar reflectivity at 3 km to ensure consistency between the two data sources. Moreover, the gridded radar-reflectivity data, based on the operational radar measurements, were quantitatively compared to the reflectivity profiles derived from the Tropical Rainfall Measuring Mission (TRMM) Precipitation Radar observations (see TRMM 2A25; [Iguchi et al., 2000]) at essentially the same times and locations. On 5, 8, 9, and 19 July 2007, the TRMM satellite observed portions of precipitating systems over the Huai River basin eight times. The horizontal distributions and magnitudes of radar reflectivity are nearly identical between the two data sources at 2–4.5 km altitude (examples are shown in Figure 2), although relatively larger discrepancies exist above 5 km, mainly because of the coarse vertical resolutions of the ground-based radar scans at the higher altitudes. Horizontal coverage of precipitation seen by the two data sets is consistent with each other at heights from ~1.5–15 km.

3. Case Description

Along with the northward shift of the subtropical high in the western Pacific near the end of June 2007, the Mei-Yu front formed over the Huai River valley, where it persisted for about two weeks and triggered the worst flood events in the Huai River valley since 1954 [Tao et al., 2008]. The MCSs occurred along the Mei-Yu front, causing heavy precipitation in the warm and moist region adjacent to the southern edge of the front. The most intense precipitation occurred on 7–8 July, with the maximum daily precipitation exceeding 200 mm. On 9 July, the water level of the Wangjiaba reservoir (location shown in Figure 1a) was up to 29.3 m, which exceeded the reservoir warning level by 1.8 m.
and prompted the Chinese Ministry of Water Resources to make a sluice in the reservoir.

The large-scale meteorological conditions during the analysis period are presented with the NCEP FNL data at 0000 UTC 8 July 2007 (Figure 3). The field of air flow within the lower troposphere was characterized by the southwesterly low-level jet along the northwestern side of the subtropical high over the western Pacific. In front of the strongest southwesterly wind center, a nearly west–east-oriented line of horizontal wind shear formed at 33°–34°N from near the surface up to 700 hPa. This shear line was colocated with the Mei-Yu front. Along the shear line, there were upward motion and the convergence of mass and water vapor, resulting in a thick layer of moist air. These large-scale features are typical of the Mei-Yu conditions in mainland China.

Horizontal distributions of radar reflectivity at 3 km with high temporal resolution clearly reveal an evolution of the MCSs (Figure 4). During the development stage of the MCSs (1800–2200 UTC 7 July), newer convection occurred along the line of horizontal wind shear at the ~850 hPa vertical layer and on the western edge of the convective lines which formed within a couple of hours. The convective cells progressed through a period of rapid growth when propagating eastward, producing stratiform precipitation to the northeast. At its mature stage (2300 UTC 7 July to 0200 UTC 8 July), the observed convective system consisted of west-east-oriented leading convective lines with stratiform cloudy regions trailing to the north and east. While the storm extended hundreds of kilometers from west to east, the widths of the convective and stratiform raining regions were only about 10–20 km and 100–200 km, respectively. After 0300 UTC 8 July, the convective elements weakened and the west-east-oriented convective line broke, while the stratiform region remained quite extensive until 0600 UTC 8 July. During the mature and dissipating stages of the linear MCSs, there was a transition zone with lower radar reflectivity located between the convective lines and the trailing stratiform regions, which is somewhat similar to the structure often broadly observed in mature squall lines [Houze, 1989]. After 0400 UTC 8 July, multiple new convective lines oriented from northeast to southwest appeared, influenced by the formation of a low-pressure system on the Mei-Yu front.

4. Design of Simulations and Analysis Method

A series of numerical simulations of the precipitation event described above were conducted using the Advanced Research WRF model, version 3.0.1, a quasi-compressible, nonhydrostatic, three-dimensional mesoscale model. Figure 5 shows the coarse-mesh and fine-mesh domains used for the simulations. The outer domain consisted of 154 by 136 grid points in x, y with a grid spacing of 30 km. The number of grid points in the inner domains were 252 by 222 (10 km) and 402 by 354 (3.3 km). All of the domains had one-way nesting. There were 30 uneven vertical levels extending from the surface to about 20 km. All simulations started at 0600 UTC 7 July and were run for 24 hr of physical time. Temperature, humidity, geopotential height, and winds from...
Figure 4. Longitude-latitude distributions of radar reflectivity (in dBZ) at 3 km altitude at ~1 hr intervals based on ground-based radar measurements obtained from 1900 UTC 7 July to 0600 UTC 8 July 2007.
Figure 5. Domain used in the Weather Research and Forecasting (WRF) simulations. Outer grid has horizontal resolution of 30 km, inner grids have resolutions of 10 km and 3.3 km, respectively. All domains have one-way nesting. Shadings represent the analysis region. White dots represent locations of the operational S-band radars. Dotted lines represent Huang River and Yangtze River.

the NCEP FNL data on 1° by 1° grids were used for the initial and boundary conditions. Analysis products were interpolated to model grid points and vertical levels using the WRF Preprocessing System. The Kain-Fritsch convective scheme [Kain, 2004] was used in the two coarser meshes, but was excluded in the inner 3.3 km domain. Other physics options used include the Mellor-Yamada-Janjic planetary boundary layer scheme [Janjic, 1990, 2002], the thermal diffusion land surface scheme [Dudhia, 1996], the Rapid Radiative Transfer Model (RRTM) longwave radiation [Mlawer et al., 1997], and the Dudhia shortwave radiation scheme [Dudhia, 1989].

Simulations were conducted with varying microphysical parameterizations to investigate their roles in simulating the structure and evolution of the MCSs, with a focus on the convective-stratiform precipitation processes. Table 1 summarizes the simulations. Three simulations (2M, 1M-T, 1M-L) were performed with the microphysics schemes available in WRF, namely, the Morrison, Thompson, and Lin schemes, respectively. The three cloud microphysics schemes are bulk microphysics parameterizations (BMPs). Hydrometeors are separated into five species (cloud water, rain, cloud ice, snow, and graupel) in the schemes. The Morrison scheme predicts the mass mixing ratios and number concentrations of cloud ice, snow, rain, and graupel, and the mixing ratio of cloud water [Morrison et al., 2009]. The Thompson scheme predicts the mixing ratios of the five hydrometeor species and the number concentration of cloud ice, incorporating a number of improvements to physical processes and computer coding compared to earlier single-moment BMPs [Thompson et al., 2008]. The Lin scheme, developed at Purdue University [Chen and Sun, 2002], is a single-moment BMP based on Lin et al. [1983] and Rutledge and Hobbs [1984], with modifications for water-ice saturation adjustment to calculate cloud water condensation and cloud ice deposition (following [Lord et al., 1984] and [Tao et al., 1989]) and ice sedimentation.

The results from 2M and 1M-L differ the most significantly, and the deficiencies in the 1M-T results appear to be closely associated with the intercept parameter of rain size distribution ($N_{0r}$). To better understand the results from 2M, 1M-L and 1M-T, five experimental simulations were conducted with modifications to the standard Morrison and Thompson schemes, respectively. In the 2M-mod, 2M-Vg, 2M-sat, and 2M-Nr simulations, some aspects of the Morrison scheme were modified using the Lin methods to replace the Morrison methods; otherwise they were identical to 2M. The 2M-mod simulation included modifications for three aspects of the microphysics parameterization: (1) calculation of cloud ice deposition and cloud water condensation at temperatures below 0°C, (2) calculation of graupel fall speeds, and (3) determination of $N_{0r}$. The modifications for (1), (2) and (3) were applied, respectively, to 2M-sat, 2M-Vg, and 2M-Nr. The last simulation, 1M-T-Nr, was identical to 1M-T, except for that both rain number concentration ($N_r$) and rain mixing ratio ($q_r$) are predictive variables in 1M-T-Nr using the Thompson scheme release with WRF version 3.1, whereas 1M-T diagnoses $N_r$ [Thompson et al., 2008]. Detailed descriptions of (1) and (2) and of the diagnostic calculation of $N_{0r}$ in 1M-T are given in Appendix A.

The Morrison, Thompson, and Lin schemes represent rain size distribution by a gamma function:

$$N(D) = N_{0r}D^{\mu_r}e^{-\lambda_D D}$$

where $N_{0r}$, $\mu_r$, and $\lambda_D$ are the intercept, shape, and slope parameters of the rain size distribution, and $D$ is the particle diameter. In the simulations, $\mu_r = 0$ was used. Thus, the size distributions for rain are inverse exponential functions. In 2M, 2M-Vg, 2M-sat, and 1M-T-Nr, $N_{0r}$ was derived from the predicted $q_r$ and $N_r$ and the specified $\mu_r$. That is, these

<table>
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<tr>
<td>2M</td>
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<tr>
<td>1M-T</td>
<td>Standard Thompson scheme</td>
</tr>
<tr>
<td>1M-L</td>
<td>Standard Lin scheme</td>
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<tr>
<td>2M-mod</td>
<td>Morrison scheme, except that (a) calculations of cloud ice deposition and cloud water condensation at temperatures below 0°C, (b) calculation of graupel fall speed, and, (c) specification of $N_{0r}$ are the same as Lin</td>
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<tr>
<td>2M-Vg</td>
<td>Morrison scheme, except that calculation of graupel fall speed is the same as Lin</td>
</tr>
<tr>
<td>2M-sat</td>
<td>Morrison scheme, except that calculations of cloud ice deposition and cloud water condensation at temperatures below 0°C are the same as Lin</td>
</tr>
<tr>
<td>2M-Nr</td>
<td>Morrison scheme, except that specification of $N_{0r}$ is the same as Lin</td>
</tr>
<tr>
<td>1M-T-Nr</td>
<td>Thompson scheme, except that $N_r$ is predicted</td>
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*All simulations had 30 half levels and 3.3 km horizontal resolution within the fine D3 domain. No convective parameterization scheme was used in domain D3 and the Kain-Fritsch scheme was used for domains D1 and D2.
simulations used a two-moment approach for rain prediction. All other simulations utilized a one-moment approach for rain prediction. They predict \( q_r \) and diagnose \( N_r \) with \( N_{0r} \), either predetermined as a constant value (8 × 10^4 m^{-2} in 1M-L, 2M-Nr, and 2M-mod) or empirically diagnosed (in 1M-T).

[17] Rain flux \( F_r \) can be expressed as \( F_r = \rho q_r V_r \), where \( \rho \) is air density, \( q_r \) is rain mixing ratio, and \( V_r \) is mass-weighted rain fall speed. Below the freezing level, rain evaporation is the main process that decreases \( q_r \). The rain evaporation rate is essentially calculated according to the concepts of diffusional growth originally developed by Byers [1965]. Specifically, the Thompson scheme follows Srivastava and Coen [1992, equation 16]; both the Morrison scheme [Morrison et al., 2009, equation 4] and the Lin scheme [Lin et al., 1983, equation 52] use an approach similar to Orville and Kopp [1977], Rutledge and Hobbs [1983], and Reisner et al. [1998]. For a given \( q_r \), \( N_{0r} \) is the only size distribution parameter that impacts not only the rain evaporation rate but also \( V_r \) in the schemes, since \( \lambda_r \) can be expressed in terms of \( N_{0r} \) and \( q_r \). An increase in \( N_{0r} \) for a given \( q_r \) means that rain size distribution shifts toward small–size rain drops. Therefore, larger \( N_{0r} \) leads to stronger rain evaporation for given \( q_r \) and environmental conditions and smaller \( V_r \) for a given \( q_r \), both contributing to a reduction of \( F_r \). An increase in \( N_{0r} \) for a given \( q_r \) also means a reduction in radar reflectivity.

[18] In order to analyze the convective–stratiform precipitation processes in the simulations, we separate the 3-D model domain into clear sky and cloudy regions, the latter of which consists of deep convective, shallow convective, and stratiform cloudy columns, using a partitioning method that is similar to Xu [1995] and based largely on observations of the kinematic structures of MCSs [Houze, 1977, 1993; Zipser, 1977]. A grid column is clear sky if the vertically integrated mass of cloud (cloud water, cloud ice) and precipitation (rain, snow, graupel) in the column is less than 0.2 kg m^{-2}. A deep convective region consists of a convective “core” and four adjacent grid columns (one grid column is 3.3 by 3.3 km²). A core contains at least one deep convective grid column. The horizontal distribution of the maximum cloud draft strength below the melting level \( (w_{\text{max}}) \) in a grid column is used as the primary variable to find deep convective columns. A grid column is defined as “deep convective core” if it satisfies at least one of the following three criteria: (1) \( w_{\text{max}} > 5 \text{ m s}^{-1} \), (2) \( |w_{\text{max}}| \geq 2 \times |w_{\text{max}}|_{\text{d}} \), where \( w_{\text{max}} \) is \( w_{\text{max}} \) averaged at the four adjacent grid columns, or (3) the surface precipitation rate greater than 25 mm h^{-1}. A grid column of shallow convection must have positive layer-mean (ground to 0°C level) vertical air velocity, rainwater path at heights below the 0°C level less than 0.1 kg m^{-2}, cloud water path at heights below the 0°C level greater than 0.4 kg m^{-2}, and cloud ice path less than 0.2 kg m^{-2}. Grid columns that are not “clear sky”, “deep convection,” or “shallow convection” belong to the stratiform region. Lang et al. [2003] compared six different convective–stratiform separation techniques, including Xu [1995], and found that they produced results that qualitatively agreed. However, the quantitative differences were significant.

[19] Although most fields are directly output by WRF, the radar reflectivity factor, \( Z_e \), must be calculated. In this study, \( Z_e \) is calculated from integration of the size distributions for rain, snow, and graupel following Smith [1984]. Since we are primarily interested in comparisons to the available radar data, only Rayleigh scattering is considered whereas Mie scattering is ignored. This assumption is justified for the relatively large wavelength used in this study (10 cm). For frozen species, a prefactor is used to compensate for the fact that the dielectric factor applies to water and not ice. The special case of partially melted snow and graupel utilizes the code of Ulrich Blahak [Blahak, 2007] that allows for different ice lattice and water coating assumptions. This produces a radar bright band that appears physically reasonable and improves upon the assumption of no meltwater.

5. Evaluation of the Simulated Precipitation

[20] The 2M, 1M-T, and 1M-L simulations produced nearly west–east oriented rain bands at the surface, as was observed (Figure 6). However, the rain bands in 1M-T and 1M-L are narrower than observed, lacking the stratiform precipitation to the north of the convective line. They also overestimated the maxima of accumulated precipitation. In contrast, the 2M simulation produced a broader stratiform rain region. However, it underestimated the magnitude of the maximum surface precipitation.

[21] Comparing the time series of horizontal–averaged, hourly accumulated rainfall rate during 1700 UTC 7 July to 0600 UTC 8 July (Figure 7), one can see that the simulations, especially 1M-L and 1M-T, underestimated the rainfall rates during the MCSs’ development–to–mature stage (before 0000 UTC 8 July), and all simulations underestimated the rainfall rates during the MCSs’ mature–to–dissipation stage (after 0000 UTC 8 July), when stratiform precipitation contributed more significantly to the total rainfall (Figure 4).

[22] To compare the vertical distributions of the observed and simulated radar echo during the MCSs’ mature stage (2300 UTC 7 July to 0200 UTC 8 July), cross sections of the observed and simulated radar reflectivity were sampled every 6 min and 5 min, respectively. At a given time, we chose multiple cross sections between 115°E and 118°E that were nearly perpendicular to the convective line and at an even distance of ∼10 km from each other. The instantaneous cross sections at the same time were first averaged as a function of the relative distance from the leading edge of the convective core and then temporally averaged. The line- and temporal-averaged cross sections (Figure 8) exhibit notable differences in the fine-scale structures of radar reflectivity between the observation and simulations also among the simulations. Note that radar reflectivity near the surface was missed by the observations because of the limit of the radar scan (Figure 8a). In the 2M simulation (Figure 8b), large reflectivity (40 dBZ) within the deep convective cores extended to about 6 km, which is almost the same distance as occurred within the observations. However, the 2M simulation underestimated radar reflectivity in the stratiform region at heights of 5–10 km and the width of the stratiform precipitation band by 20–30 km. The 1M-T (Figure 8c) produced an extensive stratiform region above the 0°C level (∼5.2 km), but the stratiform region decreased too rapidly toward the surface, suggesting excessive rain evaporation rates. Moreover, radar reflectivity within the stratiform region was underestimated and exhibited a thin “transition band” of smaller radar reflectivity just below the 0°C level that was not observed. The 1M-L (Figure 8d) produced the smallest area of strat-
form precipitation, not only at the surface but also aloft. The 40 dBZ echo in the convective cores from 1M-T and 1M-L extended up to ∼8 km in altitude, which is higher than the ground radar observation and the 2M simulation, suggesting more intense convection in 1M-T and 1M-L, which is consistent with their larger amounts of surface precipitation (Figures 6 and 7).

[23] Distributions of the frequency of occurrence of Ze (in dBZ) at 3 km from the 1M-T and 1M-L simulations, normalized by the number of total samples, are flatter than those from the radar observations and the 2M simulation (Figure 9). That is, 1M-T and 1M-L produced relatively more occurrences of Ze in the two tails of the distributions (i.e., greater than 50 dBZ and less than 25 dBZ, respectively), and underestimated the occurrence of Ze in the middle range.

[24] In summary, in terms of the horizontal distribution of the 13 h accumulated surface precipitation amount, the temporal evolution of domain-averaged surface rainfall rates, and the histogram of reflectivity at low altitudes (such as 3 km), the results from the 2M simulation are closer to the observations than the 1M-L and 1M-T simulations. The 1M-L and 1M-T simulations overestimate convective precipitation and underestimate stratiform precipitation at the surface. However, all of the simulations overestimate surface rainfall rates during the MCSs’ development-to-mature stage and underestimate surface rainfall rates during the MCSs’ mature-to-dissipation stage. Moreover, substantial differences in the radar reflectivity structure of the MCSs are found between each simulation and the observations. The differences between the simulations and the observations may be caused by multiple factors, such as deficiencies associated with the physical schemes for other processes (e.g., turbulence, land surface) and uncertainties in the initial conditions, whereas the simulated radar reflectivity structure is significantly impacted by the parameterization of cloud microphysics.

6. Stratiform-Convective Precipitation Processes in the Simulations

[25] Time series of the domain-averaged surface rainfall rates and rain area fractions at 5 min intervals from 1700 UTC 7 July to 0600 UTC 8 July in the deep convective, shallow convective, and stratiform regions are first compared among the 2M, 1M-T, and 1M-L simulations. The averages in the shallow convective regions are negligible compared to those in the deep convective and stratiform regions and are thus not shown. Figure 10 clearly exhibits that 1M-L and 1M-T produced more intense convective rainfall and a slightly larger area of convective regions than 2M, which produced greater stratiform rainfall rates and...
broader stratiform regions during the period. The 1M-L simulation generated the largest convective rainfall rates and the smallest area of stratiform precipitation during most of the period.

6.1. Processes in the Stratiform Regions

Previous studies have suggested the important impact of ice microphysics [Fovell and Ogura, 1988; McCumber et al., 1991] and rain evaporation [Li et al., 2009; Morrison et al., 2009] on simulations of stratiform precipitation. To explore why the 2M simulation produced a significantly larger amount and more extensive area of stratiform precipitation than 1M-T and 1M-L, this section compares three processes in the stratiform regions among the three simulations, including the increase of the ice-phase hydrometeor (cloud ice, snow, and graupel) mixing ratio \((q_i + q_s + q_g)\) caused by detrainment from the deep convective regions, depositional growth of hydrometeors, and rain evaporation in the stratiform regions. The results are shown in Figures 11 and 12.

The source of \(q_i + q_s + q_g\) in the stratiform regions caused by horizontal advection from the deep convective regions (Figure 11a) was negligible in 1M-L, but considerably larger in 1M-T and 2M, being more (less) significant at 10–15 km (6–10 km) in 1M-T than 2M. Depositional growth of hydrometeors in the stratiform regions (Figure 11b) was the largest in 2M, consistent with the strongest mesoscale upward motion at the upper levels (7–12 km) of the stratiform regions. The amount of \(q_i + q_s + q_g\) was the least in 1M-L (Figure 11c), consistent with the negligible horizontal transport of hydrometeors from the deep convective regions and resulting in the least amount of rain production by melting of snow and graupel (Figure 12b) and the smallest rain flux (Figure 12d). The 2M and 1M-T simulations produced comparable amounts of \(q_i + q_s + q_g\) in the stratiform regions, but the simulations peaked at 12 km (1M-T) and 8 km (2M), respectively, consistent with their different altitudes of detrainment (Figure 11a).

Among the 2M, 1M-T and 1M-L simulations, rain evaporation in the stratiform regions was greatest in 1M-T (Figure 12a), resulting in the most rapid decrease of \(q_i\) toward the surface in 1M-T (Figure 12b). At heights between 2 km and 5 km, the mean \(q_i\) in 1M-T has about the same magnitude as that in 2M (Figure 12b). However, the mean \(V_r\) at these heights is smaller in 1M-T than in 2M (0.7–1.0 m s\(^{-1}\) versus 1.4 m s\(^{-1}\), Figure 12c). The stronger rain evaporation and smaller \(V_r\) in the 1M-T stratiform regions are closely related to the diagnostic method used to determine \(N_{0r}\) in the Thompson scheme, which results in an overestimate of \(N_{0r}\) in the stratiform regions in 1M-T relative to 2M (and 1M-L). The mean \(V_r\) are a few times smaller in 1M-T than 2M, with somewhat larger differences at heights below 2 km than at 2–5 km (Figure 12c). This is caused by both smaller \(q_i\) (Figure 12b) and larger \(N_{0r}\) in 1M-T. As a result, rain flux in 1M-T is ∼90% of 2M at ∼4 km, whereas at the surface it is only 60% of 2M. The overestimate of \(N_{0r}\) also contributes to the smaller \(Z_r\) in the stratiform region of 1M-T (Figure 8c).

Therefore, the lack of stratiform precipitation at the surface in 1M-T is closely related to large values of \(N_{0r}\) in the stratiform regions which impact the strength of rain evaporation and rain fall speed, and thus rain flux, whereas in 1M-L the lack of stratiform precipitation at the surface is closely associated with insufficient amounts of ice-phase hydrometeors and positively buoyant air which have been detrained from the deep convective regions to the stratiform regions.

6.2. Processes in the Deep Convective Regions

Why do the amounts and heights of detrainment from the deep convective regions to the stratiform regions differ so significantly among the simulations? Because of mass continuity (i.e., slowing of vertical air velocity with height implies divergence of air mass flux in the horizontal), the heights where detrainment occurs are closely related to the vertical distribution of vertical air velocity \((w)\) in the deep convective regions. Vertical air velocity is related to buoyancy, which is determined in part by latent heating and condensate loading. In addition to the detrainment of hydrometeors, we also examine the detrainment of buoyancy from the deep convective regions, since this can influence the strength of mesoscale upward motion and hence the depositional growth of ice-phase hydrometeors at the upper levels of the stratiform regions. To better understand the differences among the simulations in terms of the source of \(q_i + q_s + q_g\) by detrainment from the deep convective regions and by depositional growth at upper levels of the stratiform regions, \(w, q_i + q_s + q_g,\)
Figure 8. Line- and temporal-averaged vertical cross sections of radar reflectivity from (a) observations and simulations (b) 2M, (c) 1M-T, and (d) 1M-L. The averaging method is described in the text.

Figure 9. Distributions of normalized occurrence frequency of radar reflectivity at 3 km from observations (grey solid curve) and simulations 2M (dark solid curve), 1M-T (long-dashed curve), and 1M-L (short-dashed curve).
buoyancy, and the net latent heating rate within the deep convective regions are examined (Figure 13).

In 2M, the averaged profile of $w$ in the deep convective regions peaks at 5–6 km and decreases upward and downward (Figure 13a), suggesting a significant divergence of air horizontally at the midlevels (6–10 km) where large amounts of ice-phase hydrometeors occur (Figure 11b) and the air is the most positively buoyant (Figure 13c). In 1M-T, the mean $w$ profile peaks at 6 km, as in 2M; it decreases, however, with increasing height at a slower (faster) rate from 6 to 11 km (11–15 km) than in 2M, indicating less (more) horizontal divergence in 1M-T at these heights. Combined with the peak of $q_i + q_s + q_g$ at 12 km within the deep convective regions (Figure 11c), the strong horizontal divergence at 11–15 km

**Figure 10.** (a, b) Temporal evolution of domain-averaged rainfall rate and (c, d) rain area fraction at 5 min intervals during 1700 UTC 7 July to 0600 UTC 8 July 2007, as obtained from simulations 2M (solid curve), 1M-T (long-dashed curve), and 1M-L (short-dashed curve). The precipitation is classified into deep convective (Figures 10a and 10c) and stratiform (Figures 10b and 10d).

**Figure 11.** Vertical profiles of temporally- and domain-averaged quantities in the stratiform regions from simulations 2M (solid curves), 1M-T (long-dashed curves), and 1M-L (short-dashed curves). (a) Change rate of ice-phase hydrometeor mixing ratio due to horizontal advection. (b) Change rate of ice-phase hydrometeor mixing ratio due to depositional growth. (c) Ice-phase hydrometeor mixing ratio. The averaging period is 1700 UTC 7–0000 UTC 8 July 2007.
in 1M–T led to the detrainment of large amounts of ice to the stratiform region at these heights. In 1M–L, the peak of the mean $w$ profile is located at 11 km, suggesting insignificant horizontal divergence at heights below 11 km, if there was any locally. These features are generally consistent with Figure 11a. Moreover, the mean values of $w$ are larger at 8–14 km in 1M–L and 1M–T than those occurring in 2M (Figure 13a), which is consistent with the larger positive buoyancy at 5–12 km (Figure 13c) that is contributed to by the greater latent heating rates at the mid- and high-levels (Figure 13d) in these runs. The larger mean values of $w$ at 8–15 km in 1M–L are also a result of less condensate loading at these heights (Figure 13b). The greater latent heating rates in 1M–L at heights of 7–12 km relative to 2M are attributed, at least partially, to the calculation of graupel fall speed ($V_g$) and combined cloud water condensation and cloud ice deposition at temperatures below 0°C (as will be shown in section 7). The larger latent heating rates in 1M–T over 2M at heights above 5 km are mainly due to differences in the condensation and deposition processes (Figure 14). However, it is very difficult to pinpoint specific aspects of the Morrison and Thompson schemes that caused the differences in condensation–deposition, since the two schemes differ significantly in many ways. For example, the Thompson scheme utilizes a unique representation for snow prediction [Thompson et al., 2008].

To investigate how microphysical processes affect vertical air motion at the grid scale, statistical distributions of updrafts and downdrafts are examined. An updraft or downdraft is defined here as any grid box where $w$ is greater than 1 m s$^{-1}$ or less than −1 m s$^{-1}$. This definition is the same as in McFarquhar et al. [2006] and similar to definitions used by Jorgensen et al. [1985], Tao et al. [1987], and Houze [1993]. Normalized frequency distributions of updrafts and downdrafts from the simulations show large differences at most heights among the simulations. Mann-Whitney U tests confirmed that there are significant differences at a 0.05 confidence level (Table 2). Compared to 1M–T and 1M–L, the 2M
simulation produced more intense updrafts at 5–8 km, weaker updrafts at 9–12 km, and weaker downdrafts at the low levels (≤4 km).

These results suggest that the various microphysics schemes significantly impact the coupling between microphysics and dynamics (e.g., the association and feedback between latent heating, condensate loading, and w) and further influence the convective rainfall intensity and formation-evolution of stratiform precipitation.

7. Results From Experimental Simulations

7.1. Comparison Among 2M-Vg, 2M-sat, 2M-Nr and 2M

As described in section 4, in the 2M-Vg, 2M-sat, and 2M-Nr simulations, the standard Morrison cloud microphysics scheme was modified using the Lin methods to replace the Morrison methods; otherwise the schemes were identical to 2M. The modification made in 2M-Vg involved the calculation of $V_i$; in 2M-sat the modification was the calculation of cloud ice deposition–cloud water condensation at temperatures below 0°C; and in 2M-Nr the modification was the determination of $N_0$, i.e., a constant $N_0$ of $8 \times 10^6$ m$^{-4}$ was used in 2M-Nr.

Stratiform precipitation at the surface was generally reduced in 2M-Vg, 2M-sat and 2M-Nr compared to 2M (Figure 15b). The difference in stratiform precipitation between 2M-sat and 2M and between 2M-Vg and 2M is closely related to coupling between microphysical and dynamical processes in the deep convective regions. Compared to 2M, the net latent heating rates were intensified at 8–12 km, with the intensification being greater in 2M-sat than 2M-Vg, but weaker at heights below 8 km in both 2M-sat and 2M-Vg (Figure 16a). These changes in net latent heating were caused by microphysical processes associated with condensation-evaporation, deposition-sublimation, freezing, and melting (Figure 17), reflecting the complicated interactions among the four types of microphysical processes.
as well as between graupel sedimentation and these processes. The ice mixing ratios \((q_i + q_s + q_g)\) were reduced, especially in 2M-Vg, relative to 2M (Figure 16b), meaning less condensate loading. Correspondingly, the convective updrafts were intensified at 8–13 km in 2M-Vg and at 8–15 km in 2M-sat (Figure 16c). In other words, using Lin methods to replace Morrison methods in the calculation of \(V_g\) in 2M-Vg and cloud water condensation–cloud ice deposition at temperatures below 0°C in 2M-sat produced vertical distributions of latent heating, \((q_i + q_s + q_g)\), and \(w\) that were closer to the 1M-L results (Figures 13d, 13b, and 13a). Consequently, the strength of the detrainment of hydrometeors (and air buoyancy) was significantly reduced at 5–10 km and slightly increased above 10 km (Figure 16d). At the same time, depositional growth in the stratiform regions was also weakened significantly (Figure 16e). As a result, \((q_i + q_s + q_g)\) aloft in the stratiform regions was reduced (Figure 16f), with the peak \(q_i + q_s + q_g\) of 0.2 g kg\(^{-1}\) in 2M (Figure 11c) reduced by \(\sim 0.05\) g kg\(^{-1}\). The smaller amount of \((q_i + q_s + q_g)\) aloft in the stratiform regions in 2M-sat and 2M-Vg resulted in less \(q_r\) formed by melting of snow and graupel (Figure 18a; the peak of \(q_r\) was only 0.032 g kg\(^{-1}\) in 2M-sat and 2M-Vg versus 0.045 g kg\(^{-1}\) in 2M) and reduced stratiform precipitation at the surface (Figures 15b and 18c).

[36] The 2M-Nr simulation produced more substantial differences from 2M compared to 2M-Vg and 2M-sat, with more significant decreases in both rainfall amount and horizontal span of the stratiform regions (Figures 15b and 15d). The vertical distributions of \(q_r\), averaged over the stratiform regions, peak at 4.3 km, where the average \(q_r\) from 2M-Nr is slightly larger than 2M (Figure 18a), indicating that rain production in the stratiform regions by snow and graupel melting was slightly larger in 2M-Nr. At the surface, however, \(q_r\) was smaller in 2M-Nr than in 2M (0.020 vs. 0.023 g kg\(^{-1}\); Figure 18a), reflecting stronger rain evaporation in 2M-Nr.

Figure 14. Comparison between the 2M (solid curve) and 1M-T (dashed curve) simulations of vertical profiles of latent heating rate in the deep convective regions by microphysical processes. (a) Condensation. (b) Deposition. (c) Freezing. (d) Melting. Latent heating rates are averaged during 1700 UTC 7 July to 0000 UTC 8 July 2007.
Furthermore, the mean $V_r$ was a few times smaller in 2M-Nr than 2M at all heights below the freezing level (Figure 18b). The greater rain evaporation and smaller $V_r$ both contributed to more rapid reduction of rain flux toward the surface and weaker surface precipitation (Figure 18c) in 2M-Nr, and these factors were closely related to the difference in $N_{or}$ between 2M and 2M-Nr. The constant value of $N_{or}$ ($8 \times 10^6 \text{ m}^{-3}$) used in 2M-Nr was generally larger than the predicted $N_{or}$ in 2M (Figure 19a), especially at altitudes below 3 km. This resulted in greater rain evaporation in 2M-Nr than 2M for a given $q_r$ and environmental conditions and smaller $V_r$ for a given $q_r$.

7.2. Comparison of 2M-mod With 2M and 1M-L

[37] The modifications made to 2M-Vg, 2M-sat, and 2M-Nr were combined in 2M-mod. Compared to 2M, the most significant difference in 2M-mod is the more intense convection and the much narrower and weaker region of trailing stratiform precipitation (Figure 20). Although the convective precipitation rates from 2M-mod are still less than 1M-L, a clear trend with the 2M-mod results generally changing from the 2M results toward the 1M-L results is evident. As shown in Figures 20 and 15, the differences between 2M-mod and 2M are more significant than those between 2M-Vg and 2M, 2M-sat and 2M, and 2M-Nr and 2M.

7.3. Comparison of 1M-T-Nr With 1M-T and 2M

[38] The simulation 1M-T-Nr is identical to 1M-T, except that the two-moment approach is used to predict the rain size distribution in 1M-T-Nr and the one-moment approach is used in 1M-T with $N_{or}$ empirically diagnosed. The most distinct difference between 1M-T-Nr and 1M-T is that the stratiform precipitation at the surface during the analysis period is much heavier and more extensive in 1M-T-Nr (Figures 21b and 21d). These changes are closely related to differences in $N_{or}$ between the two simulations (Figure 19). Compared to 1M-T (Figure 19c), the $N_{or}$ predicted in the stratiform regions in 1M-T-Nr (Figure 19e) was reduced by a few orders of magnitude. The reduction of $N_{or}$ in 1M-T-Nr resulted in an increase of $V_r$, which led to larger rain flux.

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**Table 2. Mann-Whitney U Statistic and Its Significance for Convective Updrafts at Various Height Ranges and for Downdrafts at 3–4 km, As Obtained From Simulations With the Morrison, Thompson, and Lin Cloud Microphysics Schemes**

<table>
<thead>
<tr>
<th>Height (km)</th>
<th>Vertical Velocity</th>
<th>2M Versus 1M-T</th>
<th>2M Versus 1M-L</th>
<th>1M-T Versus 1M-L</th>
</tr>
</thead>
<tbody>
<tr>
<td>11–12</td>
<td>w+</td>
<td>$U$</td>
<td>$P$</td>
<td>$U$</td>
</tr>
<tr>
<td>9–10</td>
<td>w+</td>
<td>22.1812</td>
<td>0.000000</td>
<td>24.7454</td>
</tr>
<tr>
<td>7–8</td>
<td>w+</td>
<td>$-5.5747$</td>
<td>0.000035</td>
<td>$-20.5540$</td>
</tr>
<tr>
<td>5–6</td>
<td>w+</td>
<td>$-11.6286$</td>
<td>0.000000</td>
<td>$-17.5414$</td>
</tr>
<tr>
<td>3–4</td>
<td>w−</td>
<td>$-16.7640$</td>
<td>0.000000</td>
<td>$-20.0116$</td>
</tr>
</tbody>
</table>

$U$ designates Mann-Whitney U statistic; $P$ designates significance of $U$; $w^+$ equals positive vertical velocity larger than 1 m s$^{-1}$; $w^-$ equals negative vertical velocity less than $-1$ m s$^{-1}$; 2M simulation uses Morrison scheme; 1M-T simulation uses Thompson scheme; 1M-L simulation uses Lin scheme.

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**Figure 15.** Temporal evolution of (a, b) domain-averaged rainfall rate and (c, d) rain area fraction at 5 min intervals during 1700 UTC 7 July to 0600 UTC 8 July 2007, as obtained from simulations 2M (grey solid curve), 2M-Vg (dark solid curve), 2M-sat (dashed curve), 2M-Nr (dotted curve). Precipitation is classified into convective (Figures 15a and 15c) and stratiform (Figures 15b and 15d).
below 3 km (Figure 22c) even with smaller $q_r$ than 1M-T (Figure 22a). As a result, surface rain flux in the stratiform regions is greater in 1M-T-Nr than 1M-T (0.42 versus 0.29 kg m$^{-2}$ hr$^{-1}$).

The differences in the intensity of convective precipitation at the surface between 1M-T-Nr and 1M-T vary with time (Figures 21a and 21c), although the temporally averaged convective rain flux is reduced in 1M-T-Nr (Figure 22d). The temporal variations of the difference in convective precipitation amounts between 1M-T-Nr and 1M-T reflect complicated interactions between the microphysical and dynamical processes associated with the different methods to determine $N_{0r}$.

[40] Observations of raindrop size distributions in MCSs [Waldvogel, 1974; Tokay and Short, 1996; Atlas et al., 1999; Maki et al., 2001] suggest larger $N_{0r}$ for convective precipitation than for stratiform precipitation. There is significant droplet collision-coalescence in the convective regions associated with large amounts of cloud water, which produces large values of $N_r$ and correspondingly large $N_{0r}$. In contrast,
there is little droplet collision-coalescence in the stratiform regions (as rain is produced mostly by the melting of snow and graupel), resulting in relatively low $N_r$ and smaller $N_{0r}$. The 1M-T simulation produced more occurrences of large $N_{0r}$ ($>10^7$ m$^{-3}$) in the stratiform regions (Figure 19c) compared to the convective regions (Figure 19d), which is the opposite of the observed trend. With the two-moment approach used to predict rain size distribution, both 1M-T-Nr (Figures 19e and 19f) and 2M (Figures 19a and 19b) at least qualitatively reproduce the observed difference in $N_{0r}$ between stratiform and convective precipitation, which is also evident in 2M-sat and 2M-Vg (not shown).

The unrealistic transition band of smaller radar reflectivity just below the 0°C level in 1M-T (Figures 8c and 23a) is absent in 1M-T-Nr (Figure 23b). To further clarify the relationship between $Z_r$ and the method to determine $N_{0r}$, we made an offline calculation of $Z_r$ using outputs from 1M-T, but we used the constant value of $8 \times 10^6$ m$^{-4}$ to determine $N_{0r}$ instead of using the diagnostic method in 1M-T [Thompson et al., 2008]. The transition band of smaller $Z_r$ was absent in the results from this offline calculation (Figure 23c), suggesting that the transition band of smaller $Z_r$ in 1M-T was directly linked to the diagnostic method, resulting in excessive $N_{0r}$ (and thus insufficient $Z_r$) just below the 0°C level (Figures 19c and 19d).

8. Summary

Starting at the end of June 2007, a Mei-Yu front persisted over the Huai River valley in China for about two weeks and caused the worst flood events in the valley since 1954. The present study investigated convective-stratiform...
precipitation processes associated with the linear MCSs on the Mei-Yu front during a 13 hr period of 7–8 July 2007 when the most intense precipitation occurred. A total of eight simulations were performed using the Advanced Research WRF model (Table 1). These simulations were identical to each other except for differences in the cloud microphysics schemes and processes. A one-way nesting technique was applied to the domains. Results were analyzed from the inner 3.3 km domain in which parameterization for deep convection was turned off. The simulated precipitation and radar reflectivity were compared to high-quality observations of precipitation and radar reflectivity.

It was found that with the Morrison [Morrison et al., 2009], Thompson [Thompson et al., 2008], and Lin [Lin et al., 1983] schemes, the “standard” simulations (2M, 1M-T, and 1M-L) largely reproduced the observed large-scale characteristics of the precipitation event. However, notable differences were found in the morphology (i.e., radar echo structure) and evolution of the MCSs at mesoscale and cloud scale between each simulation and the observations, as well as among the simulations. All of the simulations overestimated and underestimated surface precipitation during the formation-to-mature stage and mature-to-dissipation stage, respectively, of the MCSs. Deficiencies in the simulated surface precipitation may be related to uncertainties associated with the representation of physical processes in the model (such as turbulence and land surface) and the initial conditions. Horizontal distribution of the 13 hr accumulated surface precipitation amount, temporal evolution of the domain-averaged surface rainfall rate, and normalized frequency distribution of radar reflectivity at heights of 2–4 km from the 2M simulation were closer to the observations than the 1M-L and 1M-T simulations. The 1M-L and 1M-T simulations overestimated convective precipitation and underestimated stratiform precipitation at the surface.

Analyses of the 2M, 1M-T, and 1M-L simulations as well as the five experimental sensitivity simulations suggest the following major conclusions. The more intense convective precipitation in 1M-T and 1M-L as opposed to 2M was closely associated with their greater latent heating above freezing level within the deep convective regions, which contributed to stronger updrafts above 8 km. The smaller
condensate loading at 8–15 km in the convective regions in 1M-L also contributed to stronger updrafts at these heights in 1M-L relative to 2M. The difference in latent heating between 1M-L and 2M was at least partially due to the different methods used to calculate (1) graupel fall speeds and (2) cloud water condensation–cloud ice deposition above freezing level resulting from complicated condensation–evaporation, deposition–sublimation, freezing, and melting.

Figure 20. Temporal evolution of (a, b) domain-averaged rainfall rate and (c, d) rain area fraction at 5 min intervals from 1700 UTC 7 July to 0600 UTC 8 July 2007, as obtained from the simulations 2M (dark solid curve), 2M-mod (dark-dashed curve), and 1M-L (grey solid curve). Precipitation is classified into convective (Figures 20a and 20c) and stratiform (Figures 20b and 20d).

Figure 21. Temporal evolution of (a, b) domain-averaged rainfall rate and (c, d) rain area fraction at 5 min intervals from 1700 UTC 7 July to 0600 UTC 8 July 2007 from simulations 1M-T (dark solid curve), 1M-T-Nr (dashed curve), and 2M (grey solid curve). Precipitation is classified into convective (Figures 21a and 21c) and stratiform (Figures 21b and 21d).
interactions as well as interactions between graupel sedimentation and the four processes. The stronger latent heating in 1M-T as opposed to 2M was mainly caused by the difference in the condensation and deposition processes. Moreover, when calculation of both (1) and (2) and the method to determine the intercept parameter of rain size distribution ($N_{0r}$) were all replaced in 2M with Lin methods, the simulated convective (and stratiform) precipitation processes were closer to 1M-L than the simulations with only one of these modifications.

The larger stratiform precipitation in 2M as opposed to 1M-L and 1M-T was associated with the vertical distribution of air vertical velocity in the deep convective regions, which favored detraining from the deep convective regions to occur at 5–10 km in 2M, where large amounts of ice hydrometeors and positively buoyant air were available. This influenced two important processes that impact the formation and evolution of ice hydrometeors in the stratiform regions. One process is the direct increase of ice hydrometeors in the stratiform regions by detrained ice hydrometeors from the deep convective regions. The other process is the growth of ice hydrometeors through deposition in the stratiform regions due to the link between the detrainment of buoyancy, the strength of the mesoscale updraft, and ice deposition [Morrison et al., 2009].

The underestimate of stratiform precipitation at the surface in 1M-T and 1M-L was attributed to different reasons. The underestimate was closely related to the predicted rain size distribution in 1M-T and, in contrast, associated with the detrainment process from the deep convective regions in 1M-L. In 1M-T, $N_{0r}$ was overestimated by a few orders of magnitude in stratiform precipitation relative to 2M, causing excessive rain evaporation and hence a too rapid reduction of the rain mixing ratio toward the surface. This overestimate of $N_{0r}$ also resulted in an underestimate of the mass-weighted rain fall speed. Both factors contributed to the excessive reduction of rain flux toward the surface and thus an underestimate of surface stratiform precipitation. The simulations

**Figure 22.** Vertical profiles of (a, b) rain mixing ratio and (c, d) rain flux in the stratiform regions (Figures 22a and 22c) and deep convective regions (Figures 22b and 22d), as averaged during 1700 UTC 7 July to 0600 UTC 8 July 2007 from simulations 1M-T (dark solid curve), 1M-T-Nr (long-dashed curve), and 2M (grey solid curve).
that utilize the two-moment approach to predict not only the rain mixing ratio but also the rain number concentration qualitatively reproduced the difference in \( N_0 \) between convective and stratiform precipitation, as observed in previous studies. These results are generally consistent with those based on 2-D simulations of squall lines [Morrison et al., 2009] which demonstrated the importance of \( N_0 \) representation in simulating the stratiform precipitation associated with squall lines. In 1M-L, detrainment from deep convective clouds occurred mainly at 11–15 km, where there were limited amounts of ice and positively buoyant air. The detrainment of ice hydrometeors and buoyant air from deep convective regions to stratiform regions was therefore negligible compared to the other two standard simulations, resulting in fewer ice hydrometeors being aloft and thus less rain production by the melting of snow and graupel in the stratiform regions.

Furthermore, there was an unrealistic transition band of smaller radar reflectivity just below the 0°C level in 1M-T, which was more evident in the stratiform regions than the convective regions. This unrealistic feature was directly linked to the diagnostic method used to determine \( N_0 \) [Thompson et al., 2008]. Note that the newer Thompson scheme with the two-moment approach for rain prediction is now the default Thompson scheme in WRF.

The present study focused on simulations with one model for a single heavy rainfall case. We are currently investigating the sensitivity of the results to models and cases and hope to report the results in a future publication.

Appendix A

A1. Calculation of Cloud Water Condensation and Cloud Ice Deposition

To calculate cloud water condensation–evaporation and cloud ice deposition–sublimation, the Lin scheme utilizes the water–ice saturation adjustment scheme following the approach of Lord et al. [1984] and Tao et al. [1989]. This water–ice saturation adjustment scheme requires assumptions about both the coexistence of cloud water and cloud ice at temperatures less than 0°C and the partitioning between condensation and deposition. The scheme first calculates two ratios, \( r_i \) and \( r_c \), that satisfy \( r_i + r_c = 1 \). When the cloud water mixing ratio \( q_v \) plus the cloud ice mixing ratio \( q_i \) is greater than \( 10^{-3} \) kg kg\(^{-1} \), \( r_i = q_i/(q_v + q_i) \) and \( r_c = q_v/(q_v + q_i) \); otherwise, \( r_i \) and \( r_c \) depend linearly on temperature so that \( r_i = 1 \) at \( T = 0°C \) and \( r_c = 1 \) at \( T = -35°C \). These ratios are used to calculate both the saturation vapor mixing ratio \( q^* \) and the partitioning of cloud water and cloud ice. The scheme assumes that \( q^* \) is a weighted average of the respective saturation values over liquid water \( q_{v,*} \) and ice \( q_{i,*} \) at \(-35°C \leq T \leq 0°C \). That is,

\[
q^* = q_{v,*} \times r_i + q_{i,*} \times r_c \tag{A1}
\]

If the sum of water vapor mixing ratio \( q_v \), \( q_i \), and \( q_c \) is less than \( q^* \), all cloud water is evaporated and all cloud ice is sublimated to become water vapor. Under the opposite conditions, \( q_v \) and \( q_i \) are adjusted to equal \( (q_v + q_i + q_c - q^*) \times r_i \) and \( (q_v + q_i + q_c - q^*) \times r_c \), respectively, and \( q_c \) equals \( q^* \). [51] To examine the role of the water–ice saturation adjustment scheme used in 1M-L, the experimental simulation 2M-sat is conducted, which is identical to 2M except that the water–ice saturation adjustment scheme is used. In contrast, 2M determines the deposition–sublimation of cloud ice (as well as snow and rain) using a nonsteady vapor diffusion approach, and the simulation applies a saturation adjustment approach only to cloud liquid water [Morrison et al., 2005], which is reasonable because of the short droplet phase relaxation time.
A2. Calculation of Graupel Fall Speed

[52] Calculation of the terminal velocity of graupel in the 1M-L simulation follows Chen and Sun [2002]. The mass-weighted mean terminal velocity of graupel \((V_g)\) is:

\[
V_g = \frac{\Gamma(4.5)}{6\sqrt{\rho_g}} \left(4\rho g \frac{C_D}{3C_P} \right)^{1/2} \tag{A2}
\]

where \(\Gamma\) is the gamma function, \(C_D\) (a drag coefficient) = 0.6, \(\rho_g\) is density of graupel, \(\lambda_g\) is the slope parameter of graupel size distribution, \(\rho\) is air density. In the 2M simulation, a different formula is used to compute \(V_g\):

\[
V_g = \left(\frac{\rho_{ref}}{\rho}\right)^{0.54} \frac{\Gamma(4 + b)}{6\sqrt{\rho_g}} \tag{A3}
\]

where \(\rho_{ref}\) is the reference air density (1.1 kg m\(^{-3}\)), \(a = 19.3\), and \(b = 0.37\). To examine the role of the different methods used to calculate \(V_g\) in producing the different results from 1M-L and 2M, an experimental simulation 2M-Vg is designed to be identical to 2M except equation (A2) is used to replace equation (A3). Note that equation (A2) can be written in the same form as equation (A3) with different values of \(a\) (89) and \(b\) (0.5). The larger values of \(a\) and \(b\) used with the Lin method suggest a larger fall speed of graupel for a given graupel number concentration and graupel mixing ratio in 2M-Vg as opposed to 2M.

A3. Calculation of \(N_{0r}\) in the 1M-T Simulation

[53] The diagnostic calculation of \(N_{0r}\) in the 1M-T simulation is described in detail by Thompson et al. [2008]. A concise description is given here for clarity. \(N_{0r}\) depends on not only the rain mixing ratio \((q_r)\) but also on the presence of melted snow and graupel in a column above the uppermost melting level. If there are snow and/or graupel at the first level above where melting begins and rain at the first level below where melting begins, the column is searched for the bottom of the melting layer where the snow and graupel are completely melted. Below this melting layer, the mass-weighted mean size of rain is set to the same mass-weighted mean size of melted snow or graupel (whichever is larger, but restricted to no larger than 2.5 mm median volume diameter (MVD)). At the top of the melting layer, the MVD of rain is set to the maximum computed using \(N_{0r}\) in equation (A4) and 50 \(\mu\)m.

\[
N_{0r} = \left(\frac{N_1 + N_2}{2}\right) \tanh \left[\frac{q_{a0} - q_r}{4q_{a0}}\right] + \frac{N_1 + N_2}{2} \tag{A4}
\]

where \(N_1 = 9 \times 10^9 \text{ m}^{-4}\) is an upper intercept limit, \(N_2 = 2 \times 10^6 \text{ m}^{-4}\) is a lower intercept limit, and \(q_{a0} = 1 \times 10^{-4} \text{ kg kg}^{-1}\) is essentially the transition value of \(q_r\) between the two limits. The MVD of rain is linearly increased from the top of the melting layer until it reaches the bottom of the entire melting layer. In the case of pure warm rain production, \(N_{0r}\) is determined from the predicted \(q_r\) using equation (A4). In the mixed-phase case, with both warm rain production and snow (and-or graupel) above the melting level, the larger of the computed MVD and 50 \(\mu\)m is chosen and applied to the melting level. From that level to the bottom of the melting layer, the MVD is linearly increased to the equivalent size of melted ice.

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