A Method to Merge WSR-88D Data with ARM SGP Millimeter Cloud Radar Data by Studying Deep Convective Systems

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ABSTRACT
A decade of collocated Atmospheric Radiation Measurement Program (ARM) 35-GHz Millimeter Cloud Radar (MMCR) and Weather Surveillance Radar-1988 Doppler (WSR-88D) data over the ARM Southern Great Plains (SGP) site have been collected during the period of 1997–2006. A total of 28 winter and 45 summer deep convective system (DCS) cases over the ARM SGP site have been selected for this study during the 10-yr period. For the winter cases, the MMCR reflectivity, on average, is only 0.2 dB lower than that of the WSR-88D, with a correlation coefficient of 0.85. This result indicates that the MMCR signals have not been attenuated for ice-phase convective clouds, and the MMCR reflectivity measurements agree well with the WSR-88D, regardless of their vastly different characteristics. For the summer nonprecipitating convective clouds, however, the MMCR reflectivity, on average, is 10.6 dB lower than the WSR-88D measurement, and the average differences between the two radar reflectivities are nearly constant with height above cloud base. Three lookup tables with Mie calculations have been generated for correcting the MMCR signal attenuation. After applying attenuation correction for the MMCR reflectivity measurements, the averaged difference between the two radars has been reduced to 9.1 dB. Within the common sensitivity range (210 to 20 dBZ), the mean differences for the uncorrected and corrected MMCR reflectivities have been reduced to 6.2 and 5.3 dB, respectively. The corrected MMCR reflectivities were then merged with the WSR-88D data to fill in the gaps during the heavy precipitation periods. This merged dataset provides a more complete radar reflectivity profile for studying convective systems associated with heavier precipitation than the original MMCR dataset. It also provides the intensity, duration, and frequency of the convective systems as they propagate over the ARM SGP for climate modelers. Eventually, it will be possible to improve understanding of the cloud-precipitation processes, and evaluate GCM predictions using the long-term merged dataset, which could not have been done with either the MMCR or the WSR-88D dataset alone.

1. Introduction
Clouds are one of the most fundamental aspects of our climate system and have significant impacts on climate change by reflecting solar radiation back to space and absorbing longwave radiation from the surface and the atmosphere (Wielicki et al. 1996). They are also one of the most important elements in the hydrology and energy cycle of our planet, particularly in the process of producing precipitation (Del Genio et al. 2005). Yet, understanding of the interaction between clouds and the climate, namely, cloud feedback, remains the major source of uncertainty in estimating climate sensitivity with global climate models (GCMs) (Cess et al. 1996; Randall et al. 2006). The deep convection parameterization of GCMs has the greatest uncertainty because the formative and dissipative processes of convective clouds are poorly understood (Xu and Krueger 1991). A deep convective system (DCS) normally consists of two parts: the convective core (precipitation) and the stratiform region (cirrus anvil with or without precipitation). The former is important for the atmospheric hydrological cycle, whereas the latter is dominant in atmospheric radiation because of its large spatial coverage (Lin et al. 2006).

To understand the fundamental physics and the interactions between clouds and their radiative feedbacks in the atmosphere, the Atmospheric Radiation Measurement Program (ARM; Ackerman and Stokes 2003) supported by the U.S. Department of Energy (DOE) established the research site at the Southern Great Plains (SGP) in 1992. The centerpiece of the ARM cloud...
instrument array is the millimeter wavelength cloud radar (MMCR; Moran et al. 1998), which has excellent sensitivity and large dynamic range to detect small hydrometeors, such as cloud droplets and ice crystals, but the signals can be significantly attenuated during the heavy precipitation periods (Lhermitte 1990). The MMCR provides long-term continuous profiles of equivalent radar reflectivity factor and vertical velocity at the SGP site. When combined with other ARM instruments, such as the laser ceilometer, micropulse lidar, or microwave radiometer (MWR), the hydrometeor profiles of the atmosphere can be retrieved (e.g., Dong and Mace 2003; Matrosov et al. 2006). These valuable long-term datasets facilitate the progress of research on understanding cloud microphysics and their effects on climate (e.g., Clothiaux et al. 2000; Mace and Benson-Troth 2002; Dong et al. 2005, 2006).

Because the MMCR signals could be attenuated during the heavy precipitating periods, the MMCR reflectivity measurements have some gaps and discontinuities, especially when deep convective clouds pass over the MMCR. Therefore, the MMCR measurements during the precipitation periods, even for nonprecipitating DCSs with large cloud liquid water path (LWP), are questionable (Moran et al. 1998; Kollias et al. 2007; Dong et al. 2008). Dong et al. (2008) found that the ARM MMCR-derived cloud-top height is much lower (~2 km) than the Moderate Resolution Imaging Spectroradiometer (MODIS)-retrieved effective cloud height at the ARM Tropical Western Pacific (TWP) site for nonprecipitating DCSs. They suspected that some radar signals were attenuated near cloud tops because of the high LWP (~5000 g m\(^{-2}\)) measured at the ARM TWP site. The results in the Dong et al. (2008) study motivated us to investigate precipitating and nonprecipitating DCSs at the ARM SGP site. As demonstrated in Fig. 1, the MMCR signals have been severely attenuated by heavy rainfall (rain rate > 40 mm h\(^{-1}\)), and completely attenuated above 2 km between 0250 and 0300 UTC. To provide a more complete reflectivity profile, that is, to fill in the gaps of the MMCR reflectivity measurements, other types of radars designed for detecting precipitation are required.

The Weather Surveillance Radar-1988 Doppler (WSR-88D) near the SGP site is a good complementary instrument to the MMCR because it has the particular advantage of providing the vertical distribution and horizontal coverage of precipitation (Miller et al. 1998). In this study, we used the collocated WSR-88D radar volumetric scans over the SGP site and reconstructed the WSR-88D reflectivity into a time–height series that matched the MMCR data format during the period of 1997–2006. The WSR-88D data were then merged with the MMCR reflectivity (corrected for attenuation) to fill in the gaps during the heavy precipitation periods. This merged dataset provides a more complete radar reflectivity profile for convective systems associated with heavier precipitation than the original MMCR dataset. It also provides the intensity, duration, and frequency of the convective systems as they propagate over the ARM SGP for climate modelers. Eventually, we can improve our understanding of the cloud-precipitation processes, and evaluate GCM predictions using the long-term merged dataset, which could not have been done with either the MMCR or the WSR-88D dataset alone. For example, the precipitating part of DCS might be missed if using the MMCR only, whereas the cirrus anvil from the DCS might not be detected by the WSR-88D.

2. Data

a. ARM data

The ARM MMCR operates at a wavelength of 8 mm (\(K_s\) band) in a vertically pointing mode and measures continuous reflectivity profiles as clouds and hydrometeors pass over the radar field of view (Clothiaux et al. 2000). It records equivalent radar reflectivity factors (\(Z_e\)) with a 90-m vertical resolution; a total of 167 levels starting from 105 m above ground level. The MMCR uses a 3-m-diameter antenna and has a 0.2° beamwidth, which yields lateral resolution of 35 m at the height of 10 km for a vertically directed beam. The minimum detectable \(Z_e\) of MMCR are \(-55\) dBZ at 1 km and \(-35\) dBZ at 10 km (Moran et al. 1998). For cumulus clouds, the MMCR can detect \(Z_e\) up to about \(+20\) dBZ, but Mie scattering due to increasing hydrometeor size and signal attenuation make it difficult to penetrate deep convective clouds within heavy precipitation. A constant correction of 3 dB to all levels as well as near-field corrections have been added to the MMCR reflectivities (Sekelsky 2002). The MMCR has four operational modes optimized for various clouds types and runs consecutively in a 36-s cycle (Clothiaux et al. 1999). The data used in this study are 5-min averages (the Mace PI product; more information is available online at http://iop.archive.arm.gov/arm-iop/0pi-data/). The Mace PI product (Mace et al. 2006) mimics the Active Remote Sensing of Clouds (ARSLC) product (Clothiaux et al. 1999, 2000), but the key difference between the two products is how the profiles from the four operational modes are merged. For the ARSLC product, they performed interpolation to a 9-s temporal grid (the temporal spacing of the individual modes), whereas for the Mace PI product, they estimated the most reasonable measurements for a given 90-m vertical bin from one of...
the modes during the 36-s cycle and assigned the three Doppler moment measurements from that particular mode to that bin (Mace et al. 2006).

Cloud-base height is derived from a composite of the Belfort laser ceilometers, micropulse lidar (MPL), and MMCR data (CloudBaseBestEstimate; Clothiaux et al. 2000). Because the laser ceilometer and lidar are sensitive to the second moment of the scatterer size distributions of the particle, rather than the sixth moment of the MMCR, they are virtually immune to insect contamination and precipitation particles falling below cloud base. Therefore, the estimated cloud-base height from the laser ceilometer and/or lidar is used as the lowest cloud base. Cloud temperatures are estimated from a linear temporal interpolation of ARM SGP rawinsonde soundings (4 times per day). The instantaneous soundings are first degraded to a common vertical resolution of 90 m before linear interpolation. The interpolated soundings, combined with other measurements and corrections, are denoted as ARM merged soundings (Mace et al. 2006). The cloud LWP is derived from the microwave radiometer brightness temperatures measured at 23.8 and 31.4 GHz using a statistical retrieval method (Liljegren et al. 2001). The root-mean-square (RMS) accuracies of the LWP retrievals are about 20 g m$^{-2}$ and 10% for cloud LWP below and above 200 g m$^{-2}$, respectively (Dong et al. 2000; Liljegren et al. 2001). The surface precipitation rate is measured from the tipping-bucket rain gauge at the SGP Central Facility (SCF) closest to the MMCR. All ARM data used in this study were averaged to a 5-min temporal resolution (the Mace PI product).

b. WSR-88D data

The Next Generation Weather Radar (NEXRAD) WSR-88D operates at a wavelength of 10 cm (S band) and monitors its surrounding environment in a pre-programmed sequence of 360° azimuthal sweeps at

![Image](https://example.com/image.png)
various elevation angles. Thus the WSR-88D observations represent an instantaneous measurement of radar reflectivity at a given elevation angle, azimuthal angle, and range gate. With a 1° beamwidth, the WSR-88D has a spatial resolution of 1° \times 1\,\text{km} over the ARM SGP site. Each complete sequence of azimuthal scan is defined as a “volume scan.” The volume coverage patterns (VCPs) describe the different combinations of elevation angles and processing modes used for the volume scan. Depending on the operating VCPs, the sensitivities of the WSR-88D range from \( -28 \) to \( +28\,\text{dBZ} \) in clear-air mode (VCP 32) and \( +5 \) to \( +75\,\text{dBZ} \) in precipitation mode (VCP 21, 11) (for more details, see Crum et al. 1993; Klazura and Imy 1993). Deployed in 2004, the new operation mode (VCP 12) provides greater vertical resolution at lower elevation angles and completes a volume scan in 4.1 min as opposed to 5 min for VCP 11 (Brown et al. 2000).

The WSR-88D radar used in this study is located at Vance Air Force Base, Oklahoma (KVNX; 36.7408°, −98.1278°), operated by the Department of Defense (DOD). At approximately 59.3 km west of the SGP MMCR, the KVNX radar is the closest to the SGP site. The theoretical minimum detectable \( Z_e \) of the WSR-88D at this range is between \( -25 \) and \( -30\,\text{dBZ} \) for VCP 32 and from \( -20 \) to \( -15\,\text{dBZ} \) for VCP 21 (Miller et al. 1998). In this study, the level 2 data were used to reconstruct the time–height series of the WSR-88D data over the SGP site. Because of the limited available KVNX radar data archived by the DOD before 2002, a total of 28 winter [December–February (DJF)] and 45 summer [June–August (JJA)] DCS cases were collected for this study during the period of 1997–2006. Table 1 lists the number of cases under different WSR-88D VCPs, as well as their corresponding characteristics.

c. Reconstruct time–height series of WSR-88D data at the SGP site

The general approach to collocate the WSR-88D and MMCR reflectivity measurements is illustrated in Fig. 2, and the method of processing the WSR-88D data is similar to that described by Miller et al. (1998). To compare reflectivities between the two radars, WSR-88D level 2 data were used to reconstruct the reflectivity profile for the vertical column above the SGP MMCR. The portion of each WSR-88D volume scan that intersected the MMCR-sampled column was extracted and saved. The vertical resolution of the reconstructed WSR-88D reflectivity over the SGP site is determined by the VCPs used for each case. There are approximately 12 vertical levels of WSR-88D data in VCPs 11 and 12 for reconstructing the time–height series over the SGP site, whereas there are only 5 and 8 levels in VCPs 32 and 21, respectively (Fig. 3). The height of each beam that intersected the MMCR-sampled column was represented by the center of the beam.

The criteria for selecting WSR-88D reflectivities are as follows: 1) the heading and distance from the KVNX WSR-88D (36.7408°, −98.1278°) to the ARM SGP MMCR (36.6050°, −97.4850°) are calculated using the great circle navigation equation (Rinehart 2004); 2) the WSR-88D reflectivity measurements that fall within a \( ±1° \) azimuth heading and \( ±1\,\text{km} \) range gate centered on the SGP MMCR are selected; 3) if more than one reflectivity measurement is selected within that spatial domain, the one with the closest distance and heading to the SGP MMCR is used; and 4) if no signal is detected by the WSR-88D, then it is assigned as missing value.

Figure 4 shows two examples, the first (left panel) is from wintertime primarily ice-phase clouds with virtually no liquid water retrieved by the MWR, and the second (right panel) is from summertime with light precipitation measured by the surface rain gauge and high liquid water path from the MWR. As shown in both Figs. 3 and 4, the vertical resolutions of the reconstructed WSR-88D data, particularly above 6 km, are much lower than those of MMCR observations. Notice that there are almost no signal returns in the upper layers of both cases from WSR-88D data (Fig. 4), which is mainly a result of the lower sensitivity of WSR-88D for...
detecting clouds compared to the MMCR. A more detailed comparison between MMCR and WSR-88D reflectivity measurements shows that they agree well for the winter case, but the MMCR measurements for the summer case, especially for the convective portions of clouds during the periods of 0700–0800 and 1300–1500 UTC, are much lower than those of the WSR-88D. This difference is due to the vastly different nature of the two radars, such as wavelength, scanning strategy, beamwidth, resolution, and sensitivity. Also this comparison is only from two cases. To provide a statistically reliable comparison, more cases during the winter and summer seasons are needed.

3. Statistical comparison

Comparisons between the MMCR and WSR-88D observations must be conducted carefully because of the significant vertical, spatial, and temporal differences between the two observing platforms. Therefore, it is necessary to discuss these differences before proceeding to statistical comparisons of the reflectivity measurements from the two radars.

As a consequence of being located 59.3 km away from the MMCR, the spatial resolution of the WSR-88D is about 1° × 1 km over the SGP site, and its vertical resolution (∼1 km) is much coarser than that of the MMCR (90 m). To match the two datasets vertically, a Gaussian weighted average (Rinehart 2004; represented by a parabolic curve, see Fig. 5a) was applied to the MMCR reflectivities within each collocated WSR-88D beam to represent the MMCR reflectivity measurement for that volume. The horizontal resolutions are 35 and 1000 m at an altitude of 10 km for the MMCR and WSR-88D, respectively. Although this spatial difference can cause some uncertainties in the reflectivity of convective clouds for a short time period (Straka et al. 2000), a 5-min-averaged MMCR reflectivity should reduce this difference significantly.

For the temporal resolutions, the MMCR data are 5-min averages, whereas the reconstructed WSR-88D data are the instantaneous reflectivity measurements at
successive vertical levels over the SGP during a period of 4–10 min, depending on the VCP mode because the WSR-88D scans 360° at different elevation angles. It is assumed that convective cloud systems do not evolve significantly over the course of 4–10 min for the WSR-88D to complete a full volume scan. Therefore, the reconstructed WSR-88D dataset is considered to be a radar reflectivity profile comparable to that of the MMCR. Moreover, reflectivity measurements depend on absolute calibration. In this study, both radars are assumed to be correctly calibrated.

From theoretical study, there is no attenuation in detecting ice particles from both radars, but there is light attenuation for the WSR-88D measurements and severe attenuation for the MMCR measurements in detecting precipitation (Moran et al. 1998; Lhermitte 1990). To quantitatively investigate the difference in reflectivity measurements from the two radars, the cases from both winter and summer seasons have been carefully selected from the matched MMCR and WSR-88D dataset. The cases selected during winter seasons with primarily ice-phase convective clouds [e.g., Fig. 4 (left panel)] were used as a consistency check between the two radars, whereas the summer convective cases [e.g., Fig. 4 (right panel)] were used to quantify the attenuation of the MMCR measurement for nonprecipitating DCSs.

Both winter and summer convective cases in Fig. 5 were selected during the time periods with no surface precipitation (rain rate = 0) detected by the tipping-bucket rain gauge. There are three additional criteria for selecting radar reflectivities from the winter cases: 1) the cloud temperatures are lower than 273 K; 2) the MWR-retrieved LWP is less than 20 g m⁻² (considering the uncertainty of the retrieved LWP; Dong et al. 2000); and 3) the reflectivity measurements are above the cloud base. The radar reflectivities from the summer cases were selected to investigate the signal attenuation of MMCR measurement from the liquid cloud droplets rather than rain drops with the following two additional criteria: 1) the cloud temperatures are higher than 273 K and 2) the contiguous cloud tops exceed 5 km.

Figures 5b and 5c show the scatterplots of MMCR versus WSR-88D reflectivities for the winter and summer cases, respectively. For the winter cases (Fig. 5b),
the two radars provide very consistent reflectivity measurements for primarily ice-phase convective clouds, considering the large sampling differences discussed above. The MMCR reflectivity, on average, is only 0.2 dB lower than the WSR-88D measurement, with a correlation coefficient of 0.85. The selected samples show that the lower bounds for WSR-88D and MMCR are around 20 and 23 dBZ, respectively, whereas the upper bounds are about 30 and 20 dBZ, respectively. The reflectivity measurements from both radars agree better from −10 to +20 dBZ (from the MMCR measurement point of view) than other ranges because the two radars are sensitive to ice crystals and cloud droplets in this range. This comparison shows that the WSR-88D

![Diagram](image_url)
reflectivity measurements agree well with the MMCR measurements. The possibility of both radars’ calibration shifting to the same direction (high or low) cannot yet be ruled out, but this possibility is presumed to be very low.

Figure 5c shows the comparison for the pure liquid portion of summer nonprecipitating convective clouds. Averaged over all the differences between the matched two datasets, the MMCR measurement is 10.6 dB lower than the WSR-88D reflectivity. The correlation coefficient between the two datasets is 0.64, which is slightly lower than the value for ice-phase convective cases. The main differences seem to originate from 0 to +20 dBZ, where the WSR-88D reflectivity can be up to +40 dBZ, whereas the maximum MMCR reflectivity is around +20 dBZ. If we have a subset of the original two datasets within the common sensitivity range (−10 to 20 dBZ), the mean difference is reduced to 6.2 dB. There are several possible reasons to explain these differences: 1) the MMCR signals may be attenuated by the large LWP values, 2) the MMCR receiver is saturated at around +20 dBZ, 3) large drizzle particles (>2.5 mm) may cause the MMCR reflectivity to be smaller than Rayleigh assumption as a result of the Mie effect, and 4) the large sample volume difference between the two radars may also be a significant contributor.

To further investigate the differences with height, the two matched datasets were grouped as a function of height above cloud base, and the differences between WSR-88D and MMCR measurements were averaged for every 0.5 km. As shown in Figs. 5d and 5e, the reflectivity differences between the two radars are nearly constant with height for both winter and summer cases, although there are biases near cloud base for both seasons. There are two possible reasons to explain these biases. The first reason is that the biases may be caused by large droplets present near the cloud base. The second possible reason is that the WSR-88D measurements may include the ground contribution from its sidelobe. However, these biases still exist when the comparison is constrained for the cloud bases above 1 km (not shown). Thus, it is possible to rule out the ground clutter issue of the WSR-88D. The vertical differences for liquid-phase summer cases (Fig. 5c) are consistently large throughout 5 km above cloud base with an averaged difference of 10.6 dB and relatively large standard deviations (~5 dB). Therefore, a further study about the MMCR signal attenuation for the summer convective clouds is necessary.

4. Attenuation correction

a. Theoretical calculation

In this study, the attenuation of MMCR signal by liquid hydrometeors (mainly cloud droplets and drizzles) is the major concern. The specific attenuation $A$, which is a function of the extinction coefficient $\sigma_{ext}$, is calculated by integrating over the liquid hydrometeor drop size distribution (DSD) function $n(D)$ (Bringi and Chandrasekar 2001):

$$A = 4.343 \times 10^3 \int_{D_{min}}^{D_{max}} \sigma_{ext} n(D) dD,$$  

where the minimum and maximum diameter $D_{min}$ and $D_{max}$ are 1 and 5000 $\mu m$, respectively. At Ka band (8-mm wavelength), the size of liquid particles are comparable to the radar wavelength, thus the simple Rayleigh theory is not applicable for calculating $\sigma_{ext}$. Therefore, Mie theory is used in the calculation of the extinction efficiency $Q_{ext}$ (Bohren and Huffman 1983), which relates to $\sigma_{ext}$ through

$$\sigma_{ext} = Q_{ext} \pi r^2,$$  

where $r$ is the particle radius. For cloud droplets and drizzles, the hydrometers under consideration are assumed to be spherical particles. The extinction efficiency $Q_{ext}$ is a function of the radar wavelength, temperature, and hydrometeor type (in this case, liquid water). The DSD function considered for cloud droplets and drizzles is the gamma distribution (discussed more later), which can be expressed as

$$n(D) = \frac{N_i}{\Gamma(\nu)} \left(\frac{D}{D_n}\right)^{\nu-1} \exp\left(-\frac{D}{D_n}\right),$$  

where $N_i$ is the total number concentration; $D_n$ is the characteristic diameter, which can be related to the mean diameter $D_{mean}$ through $D_n = D_{mean}/\nu$; and $\nu$ is the shape parameter. The total number concentration $N_i$ can be derived through LWC as

$$N_i = \frac{LWC}{\pi \rho_i D_n^3 \Gamma(\nu)/(\nu + 3)},$$

As described in Eqs. (2)–(4), the specific attenuation (one way) of MMCR in Eq. (1) depends upon LWC, temperature, mean diameter, and shape parameter under the assumption of a gamma distribution. Sensitivity studies have been conducted to investigate the dependence of specific attenuation on each of the parameters (Fig. 6). As demonstrated in Fig. 6a, the specific attenuation increases linearly with increasing LWC for a given mean diameter and shape parameter. For example, for a fixed temperature of 273 K, the specific attenuation increases to 6 dB km$^{-1}$ as LWC increases.
from 0 to $3 \text{ g m}^{-3}$. The specific attenuation increases semilogarithmically with decreasing temperature (Fig. 6b).

For example, the specific attenuation increases from 2.2 to 3.2 dB km$^{-1}$ as temperature decreases from 300 to 240 K at a fixed LWC of $1 \text{ g m}^{-3}$. Specific attenuation increases less with mean diameter except for those larger than 300 $\mu$m as shown in Fig. 6c. Note that a broader distribution (i.e., smaller shape parameter $\nu = 2$; see Fig. 6d) can cause more attenuation than a narrow distribution. For $\nu = 1$, the gamma distribution in Eq. (3) can be simplified as

$$n(D) = \frac{N_\nu}{D_n} \exp \left( -\frac{D}{D_n} \right),$$

which is equivalent to the Marshall–Palmer (M–P) distribution (Marshall and Palmer 1948). Heymsfield and Hjelmfelt (1984) found that most of the liquid-phase convective cloud droplet diameters were below 450 $\mu$m with an exponential distribution using the aircraft in situ measurements over Oklahoma. Other studies (Chandrasekar and Bringi 1987; Ulbrich 1983) suggested that the particle sizes of convective clouds followed a gamma distribution with a shape parameter $\nu$ varying from 1.4 to 2.6 using both modeling and observations. Based on these previous studies, three lookup tables for correcting attenuation have been generated from the calculations in Fig. 6. They represent different DSDs: 1) gamma distribution with $D_{\text{mean}} = 100 \mu$m and $\nu = 1.5$ (Fig. 7a), 2) M–P distribution with $D_{\text{mean}} = 200 \mu$m and $\nu = 1.0$ (Fig. 7b), and 3) M–P distribution with $D_{\text{mean}} = 300 \mu$m and $\nu = 1.0$ (Fig. 7c). For a given DSD, the specific attenuation is only a function of LWC and temperature of the hydrometeors.

b. Applying the attenuation correction

As discussed in the previous section, temperature and LWC are required to obtain the attenuation for each
range gate of the MMCR. The temperature profiles are given by the ARM merged soundings in this study. To obtain LWC profile, MWR-retrieved LWP was used as a constraint. Investigations of some radar reflectivity profiles in the selected DCS cases (not shown) from the MMCR and WSR-88D show that the differences between the original reflectivity measurements of the two radars are the largest near cloud base and then decrease with height. If these differences are mainly caused by the liquid water attenuation to the MMCR signals, then we can assume that the LWC profile linearly decreases with height from cloud base to the level at $T = 240$ K in this study. The linearly decreased LWC profile gives a larger correction close to the cloud base and less correction to the cloud top, which makes the corrected MMCR reflectivity closer to that of the WSR-88D.

After obtaining the LWC and temperature profiles, the specific attenuation for each MMCR 90-m reflectivity measurement was then estimated using the aforementioned lookup tables. Because the radar signals travel through the clouds 2 times, the specific attenuation has been used twice for correcting MMCR reflectivities. The reflectivity attenuation at a given level is the result of the attenuation through all underlying layers. For example, the layer 3 attenuation of MMCR reflectivity is the sum of layers 1–3 because the radar signals have to go through layers 1 and 2, thus the attenuation of layers 1 and 2 should be used in calculating the layer 3 attenuation. After correcting attenuation using the three generated lookup tables, the MMCR reflectivities during summer (in Fig. 5e) were then compared with the WSR-88D reflectivities to quantify the difference between two radar measurements.

Figure 7 illustrates three DSDs and their corresponding averaged reflectivity differences between the MMCR and WSR-88D up to 5 km above cloud base (same as Fig. 5e). Compared to the original reflectivity differences (10.6 dB) in Fig. 5e, the averaged differences after correcting the MMCR reflectivity attenuation have been reduced by 0.7, 1.6, and 2.5 dB, respectively, for the assumed gamma and M–P distributions. The 1.6-dB reduction represents the averaged attenuation correction, whereas the 0.7 and 2.5 dB may be the lower and upper bounds of the correction. The averaged differences from bins 3–8 (1–4 km above cloud base) slightly decrease with height, which is expected after applying the attenuation correction. Figure 7d shows three averaged attenuation profiles for the selected cases as a function of height above cloud base, using the generated lookup tables with the estimated LWC and temperature profiles. Figure 7d has demonstrated that the attenuations from three DSDs monotonically increase from cloud base up to 8 km as a result of the assumed linearly decreased LWC profile and accumulated attenuation from cloud base upward.
Despite the effort to correct the MMCR attenuation using the methods described above, there are still the averaged differences of 8.2–10.0 dB between the two radars depending on the DSD. If we have a subset of the two datasets within the common sensitivity range (−10 to 20 dBZ), the mean differences are reduced to 5.3 dB. There are several possible reasons to explain these differences.

First, the MMCR receiver is possibly saturated at or before +20 dBZ, thus the MMCR reflectivity measurements around +20 dBZ are lower than the WSR-88D data. The MMCR data used in this study combine all four operating modes that cover hydrometeors with reflectivity range of approximately −50 to +20 dBZ (Clothiaux et al. 1999). The first three modes are more suitable in detecting nonprecipitating or weakly reflecting clouds, whereas the robust mode (mode 4) should be able to detect weakly precipitating clouds. Unfortunately, the robust mode was not set up to avoid saturation until 2004 and a 5-min average of this mode at the edge of sharp convective rain events (e.g., Fig. 1) might have a potential problem because mode 4 is less sensitive to nonprecipitating clouds than other modes (Clothiaux et al. 1999). Therefore, with the 5-min combined MMCR reflectivity used in this study, we suspect that the large difference found is because of the saturation of the MMCR receiver.

Second, the assumption of neglecting the volume differences between the two radars might partially contribute to the differences. However, in the best-case scenario of wintertime primarily ice-phase convective clouds where the MMCR is free of attenuation, the averaged difference (Fig. 5b) is 0.2 dB, and the two radar reflectivity measurements, on average, are almost the same. For the summer convective cases, such differences could be much larger when a few large raindrops were present in the large WSR-88D volume but missed by the MMCR. Furthermore, the criteria of selecting data in summer cases for comparison assume that the hydrometeors with sounding temperatures higher than 273 K are pure liquid. While this assumption is generally acceptable, there are some possibilities that melting ice particles could be present in the selected data. In such cases, not only would the WSR-88D measure much higher reflectivity as a result of higher power returned from the liquid shielded melting ice particles (often referred to as the bright band), but the attenuation of MMCR signal due to melting particles could be as much as 3 times higher than equivalent liquid hydrometers (Matrosov 2008). The net effect would enlarge the existing difference between the two radar measurements and could not be corrected by our proposed lookup tables.

Third, during the periods right after heavy precipitation (e.g., 0310 UTC in Fig. 1), it is possible to have water accumulated on the MMCR radome, which could significantly attenuate radar signals. The comparisons discussed in Figs. 5 and 7 assume that the accumulated water on the MMCR radome has been evaporated after 10 min of any measured precipitation. However, precipitation in some convective cases lasted for hours and such an assumption might be invalid. Last but not least, WSR-88D in its precipitation mode (VCPs 21, 11, and 12) has lower sensitivity than the clear-air mode (VCP 32) as mentioned in section 2. The small hydrometeors (e.g., reflectivities less than −10 dBZ in Fig. 5c) can be measured by the MMCR, but they are beyond the lower bound of WSR-88D measurement (about −10 dBZ).

5. Merging dataset and discussion

After investigating the MMCR attenuation, the corrected MMCR reflectivity measurements using M−P distribution with $D_{\text{mean}} = 200 \, \mu m$ in this study were then merged with the WSR-88D measurements. Because only one time stamp was recorded for each WSR-88D volume scan and it is usually not at the exact 5 min, the reconstructed WSR-88D vertical profile was approximated to the closest 5-min temporal resolution to match the MMCR data format in the merged dataset. During the heavy precipitation periods (rain rate $> 7 \, \text{mm h}^{-1}$) when the MMCR signals failed to penetrate the entire depth of the convective clouds, the WSR-88D data were used to fill in the gaps. When the MWR-retrieved LWPs are not reliable or there are no LWP retrievals at all, it is impossible to implement the attenuation correction scheme in this study. To avoid discontinuity during the transition periods in the merged dataset, the WSR-88D reflectivity measurements during the precipitating period $\pm 5 \, \text{min}$ were used to fill in the gaps of MMCR measurements. The vertical resolution of the WSR-88D data depends upon the VCP modes used at the time of the measurement. As a result of the much lower vertical resolution ($\sim 1 \, \text{km}$) of the WSR-88D, a distance-weighted linear interpolation was performed between any two levels of WSR-88D reflectivity measurements such that its vertical resolution was comparable to the MMCR measurements. If the MMCR detected weak reflectivities above the highest WSR-88D level, those reflectivities were then included in the merge dataset.

Two examples of the merged dataset have been illustrated in Fig. 8. The case on 17 June 2001 demonstrates the ability of the WSR-88D reflectivities to fill in the MMCR gap in the merged dataset. During the periods of heaviest precipitation ($\sim 0300 \, \text{UTC}$), the MMCR
signals were almost completely lost or attenuated; the WSR-88D reflectivities were used to fill in the gap of MMCR reflectivity measurements to produce a more complete reflectivity profile. Notice that during the period (0245 UTC) right before the sharp convective rain, the LWP increased dramatically up to 104 g m$^{-2}$. This huge LWP makes the transition from the attenuation-corrected MMCR reflectivity to the interpolated WSR-88D reflectivity smoothly. However, during the period (0315 UTC) right after the rain, there are no LWP retrievals, thus no attenuation correction method can be applied.

The case on 21 June 2004 clearly shows the advantage of the merged dataset compared with the original MMCR data. The merged dataset shows much higher reflectivities than the original MMCR measurements during the periods of high LWP, such as 0530–0700 UTC, 0900–1100 UTC, and 1230–1400 UTC, which is consistent with the attenuation-correction method described in section 4. If the cloud-base heights were not available during downtime of the laser instruments (~1000 UTC), the cloud bases were assumed to be at the lowest height of MMCR reflectivity measurement, with the constraint that either the rain rate was greater than zero or LWP was larger than 1000 g m$^{-2}$. This constraint is to ensure that the liquid cloud base is close to the surface and the attenuation-correction scheme can still be applied. The correction method, in general, is applicable to the nonprecipitating portion of DCS with the available LWP and sounding temperatures; for the weakly precipitating portion of the DCS, however, the correction method may need further improvement (e.g., between 0900 and 1100 UTC).

This merged MMCR and WSR-88D dataset provides a much wider range of hydrometeor reflectivity measurements from −50 to +60 dBZ compared to the original MMCR and WSR-88D datasets. The most obvious advantage of the merge dataset is being able to provide information that was lost during the period of heavy precipitation, when the MMCR failed to penetrate the entire DCS. The WSR-88D data were also used to quantify the MMCR attenuation during the

FIG. 8. Example cases of the merge dataset. (a),(e) MMCR reflectivity, (b),(f) reconstructed WSR-88D reflectivity, (c),(g) merged WSR-88D and attenuation-corrected MMCR reflectivity, (d),(h) LWP (blue line) and surface rain rate (black line). (a)–(d) 17 Jun 2001. (e)–(h) 21 Jun 2004.
6. Summary and concluding remarks

A total of 28 winter and 45 summer DCS cases over the ARM SGP site have been collected for this study during the period of 1997–2006. From the analysis of the merged dataset and the generated lookup tables for correcting MMCR signal attenuation, we have made the following conclusions:

1) For the winter cases, the two radars provide very consistent reflectivity measurements for primarily ice-phase convective clouds. The MMCR reflectivity, on average, is only 0.2 dB lower than the WSR-88D measurement with a correlation coefficient of 0.85 (both radars are sensitive to ice crystals). This result has indicated that the MMCR signals have not been attenuated for ice-phase convective clouds, and the WSR-88D reflectivity measurements agree well with the MMCR measurements.

2) For the summer nonprecipitating convective clouds, however, the MMCR reflectivity, on average, is 10.6 dB lower than that of the WSR-88D, and the averaged differences are nearly constant with height above cloud base. There are four reasons to explain these differences: (i) the MMCR signals may be attenuated by the large LWP values; (ii) the MMCR receiver is saturated at around +20 dBZ; (iii) large drizzle particles may cause the MMCR reflectivity to be smaller than Rayleigh assumption as a result of the Mie effect; and (iv) the large sample volume difference between the two radars may also be a significant contributor.

3) Three lookup tables (based on one gamma DSD and two M–P DSDs) with different mean diameters and shape parameters have been generated for correcting the MMCR signal attenuation. For a given DSD, the specific attenuation is a function of LWC and temperature of the hydrometeors. Attenuation correction for the MMCR reflectivities during the summer season was then applied to the original data and compared to the WSR-88D reflectivities again. The averaged difference between the two radar measurements has been reduced to 9.1 dB from the original 10.6 dB. For a subset of the original two datasets within the common sensitivity range (−10 to 20 dBZ), the mean differences for the uncorrected and corrected MMCR reflectivities have been reduced to 6.2 and 5.3 dB, respectively.

4) The merged MMCR and WSR-88D dataset provides a more complete time–height evolution of the reflectivity profile of convective systems, which can fill in the gaps from MMCR measurements during the heavy precipitating periods. It is our hope that this merged dataset can be used by modelers to improve our understanding of the life cycle of convective systems, as well as their associated water and energy budgets. Because satellite data are critical for studying large-scale spatial cloud, precipitation, and radiation properties, we will map the WSR-88D observed horizontal reflectivity on the Geostationary Operational Environmental Satellite (GOES)–MODIS-retrieved cloud-top properties to study the spatial structures and developmental stages, such as formative and dissipative processes, of convective systems in future studies. Eventually we can generate a 3D database from both vertical and spatial distributions of observations for modelers to simulate various stages of DCS and to improve cumulus cloud parameterizations.

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REFERENCES


