Vertical Structure of Precipitation and Related Microphysics Observed by NOAA Profilers and TRMM during NAME 2004

Christopher R. Williams and Allen B. White

Cooperative Institute for Research in Environmental Sciences, University of Colorado, and Physical Sciences Division, National Oceanic and Atmospheric Administration/Earth System Research Laboratory, Boulder, Colorado

Kenneth S. Gage and F. Martin Ralph

Physical Sciences Division, National Oceanic and Atmospheric Administration/Earth System Research Laboratory, Boulder, Colorado

(Manuscript received 21 October 2005, in final form 27 February 2006)

ABSTRACT

In support of the 2004 North American Monsoon Experiment (NAME) field campaign, NOAA established and maintained a field site about 100 km north of Mazatlán, Mexico, consisting of wind profilers, precipitation profilers, surface upward–downward-looking radiometers, and a 10-m meteorological tower to observe the environment within the North American monsoon. Three objectives of this NOAA project are discussed in this paper: 1) to observe the vertical structure of precipitating cloud systems as they passed over the NOAA profiler site, 2) to estimate the vertical air motion and the raindrop size distribution from near the surface to just below the melting layer, and 3) to better understand the microphysical processes associated with stratiform rain containing well-defined radar bright bands.

To provide a climatological context for the profiler observations at the field site, the profiler reflectivity distributions were compared with Tropical Rainfall Measuring Mission (TRMM) Precipitation Radar (PR) reflectivity distributions from the 2004 season over the NAME domain as well as from the 1998–2005 seasons. This analysis places the NAME 2004 observations into the context of other monsoon seasons. It also provides a basis for evaluating the representativeness of the structure of the precipitation systems sampled at this location. The number of rain events observed by the TRMM PR is dependent on geography; the land region, which includes portions of the Sierra Madre Occidental, has more events than the coast and gulf regions. Conversely, from this study it is found that the frequencies of occurrence of stratiform rain and reflectivity profiles with radar bright bands are mostly independent of region. The analysis also revealed that the reflectivity distribution at each height has more year-to-year variability than region-to-region variability. These findings suggest that in cases with a well-defined bright band, the vertical profile of the reflectivity relative to the height of the bright band is similar over the gulf, coast, and land regions.

1. Introduction

The North American Monsoon Experiment (NAME) is a process study aimed at determining the sources and limits of predictability of warm-season precipitation over North America with an emphasis on its seasonal to interannual variability (NAME Project Science Team 2006). Two of the overarching goals of NAME are to observe the North American monsoon system and its variability in relation to the seasonal and annual cycle of the coupled land surface–atmosphere–ocean system, and to understand the physical processes of the North American monsoon system that must be parameterized in order to improve dynamical models of the monsoon (Higgins et al. 2006). Some of the scientific objectives of NAME are to promote a better understanding and more realistic simulation of the warm-season convective processes in complex terrain, to simulate the intraseasonal variability of the North American monsoon system (NAMS), and to simulate the response of the warm-season atmospheric circulation and precipitation patterns to slowly varying, potentially predictable oceanic and continental surface boundary conditions.

To improve our understanding of the monsoon and to improve the model simulations, the NAME 2004
field campaign was held from June through September 2004 with special observations made at spatial scales from several meters to several 1000s of kilometers. The NOAA/Aeronomy Laboratory (NOAA/AL) and NOAA/Environmental Technology Laboratory (NOAA/ETL)\(^1\) collaborated during the NAME 2004 Field Campaign to operate a field site containing wind-profiling radars, precipitation-profiling radars, surface upward–downward-looking radiation instruments, and surface meteorological instruments. The observations that were collected at this NOAA profiler site and their analysis can be found online (http://www.etl.noaa.gov/programs/2004/name/).

While a single field site observing the precipitation systems that pass directly overhead cannot address all of the goals and scientific objectives of NAME, a single site can help improve our understanding of the physical processes of the monsoon system, which should lead to better precipitation parameterizations. The four main objectives of deploying profilers in support of NAME were to 1) observe the vertical structure of precipitating cloud systems, 2) estimate the vertical profile of raindrop size distributions and the vertical air motions during precipitating events, 3) better understand the microphysical processes associated with oceanic and continental rain regimes, and 4) incorporate these observations into precipitation modeling activities. This paper addresses the first three objectives of this project with the modeling activities postponed to a later study.

The first objective of the field phase of this project was to observe the vertical structure of the precipitation. Section 2 contains a description of the NOAA profilers that were deployed during NAME 2004 in order to achieve this objective. Illustrative observations are presented in section 3 to characterize the vertical structure of the precipitation, which must be known to achieve a better understanding of the microphysical processes of precipitating cloud systems.

The second objective of this project was to estimate the vertical air motion and the raindrop size distribution (DSD) as a function of altitude. Section 4 contains a description of how these estimates are made using two profiling radars operating at different frequencies. One profiler is sensitive to the Bragg scattering from inhomogeneities in the turbulent refractive index and the other profiler is sensitive to the Rayleigh scattering from hydrometeors (Gage et al. 1999). This method of using two collocated radars to estimate the vertical air motion and the DSD can only be performed by a few research groups due to the expense of operating multiple-frequency radars at the same field site.

It is important to place a single-season field campaign in a climatological context to see whether its sample is representative. To address whether the NAME 2004 profiler observations were representative, the profiler reflectivity distributions are compared with the reflectivity distributions observed by the TRMM Precipitation Radar (PR) for the 2004 season and for the 1998–2005 seasons. While the low probability of a ground-based profiler and the satellite PR simultaneously observing the same rain event prevents a direct comparison between the two instruments, a statistical comparison can be made to examine whether the field campaign observations obtained representative samples. This comparative analysis is presented in sections 5 and 6 with concluding remarks in section 7.

2. NOAA profiler deployments during NAME 2004

In support of the North American Monsoon Experiment (NAME), the NOAA/Aeronomy Laboratory and NOAA/Environmental Technology Laboratory deployed several radar profilers and surface-based instruments at one location to observe the vertical structure of the precipitation that passed directly over the NOAA profiler site. The profiler site was located at 24.28°N, 107.16°W, about 100 km north of Mazatlán near the city of Estación Obispo, in the state of Sonora, on the northwest coast of mainland Mexico. This site is about 45 km north of the location of the National Center for Atmospheric Research (NCAR) S-band dual-polarization Doppler radar (S-Pol) during the NAME 2004 campaign. The map in Fig. 1 shows the location of the profiler site relative to Mazatlán and the S-Pol scanning radar.

During the 2-month campaign from 30 July through 18 September 2004, the NOAA site contained three profiling radars, a ceilometer for measuring cloud base, a surface rain gauge, surface flux instruments, and a 10-m tower equipped with meteorological instruments. Table 1 lists the instruments used during the NAME 2004 campaign.

The profiling radars are ideal instruments for examining the vertical structure of precipitating cloud systems (Carter et al. 1995; Gage et al. 1994, 2002; Ecklund et al. 1999; White et al. 2003; Neiman et al. 2005). Figure 2 shows the antenna systems of the three NOAA profiling radars. The 915-MHz profiling radar consisted of three separate antennas enclosed in telescoping clut-

\(^{1}\) Since October 2005, NOAA/AL and NOAA/ETL were merged into the NOAA/Earth System Research Laboratory (NOAA/ESRL) and personnel who participated in the NAME 2004 field campaign are now members of the Physical Sciences Division.
ter screens with each antenna pointed in a different direction. The 449-MHz profiling radar antenna was composed of dipole elements encased in fiberglass tubes to form a collinear coaxial antenna. The 2875-MHz precipitation profiler used a fixed-dish antenna to continuously direct its radar beam vertically.

The three profiling radars operated at three different frequencies to observe different attributes of the atmosphere directly overhead. As the operating frequency increases from 449 to 2875 MHz, the radar becomes less sensitive to the backscattering of energy from the irregularities in the refractive index due to turbulence (Bragg scattering) and more sensitive to the backscattering of energy from distributed targets (Rayleigh scattering). See Ralph (1995) and Gage et al. (1999) for studies that examine the sensitivity of the Bragg and Rayleigh scattering processes as a function of operating frequency. We use only the 449- and 2875-MHz profilers here to exploit the advantages of these relatively low and high frequency radars to measure air motions and precipitation, respectively. The observations collected by the 915-MHz profiler, which focused on wind profiling, and the other instruments at the NOAA profiler site are described in detail in Hartten et al. (2007, manuscript submitted to J. Climate).

Table 1. Instruments at the NOAA profiler site near Estación Obispo, Sonora, Mexico.

<table>
<thead>
<tr>
<th>Instrument</th>
<th>Purpose</th>
</tr>
</thead>
<tbody>
<tr>
<td>2875-MHz vertically pointing profiler</td>
<td>Vertical structure of precipitation</td>
</tr>
<tr>
<td>449-MHz vertical air motion and wind profiler</td>
<td>Estimate the vertical air motion during precipitation and the horizontal wind outside of precipitation</td>
</tr>
<tr>
<td>915-MHz wind profiler</td>
<td>Estimate the horizontal wind up to 5-km altitude</td>
</tr>
<tr>
<td>Cloud boundary ceilometer</td>
<td>Estimate altitude of cloud base</td>
</tr>
<tr>
<td>Surface flux array</td>
<td>Estimate the upward and downward radiation fluxes</td>
</tr>
<tr>
<td>10-m meteorological tower</td>
<td>Temperature, humidity, and pressure measurements at 2 and 10 m</td>
</tr>
</tbody>
</table>

Fig. 1. Map showing location of the NOAA profiler site relative to the NCAR S-Pol scanning radar and the city of Mazatlán. The circles indicate the 150-km range rings around the scanning radars.

a. Operating modes: 2875-MHz profiling radar

The main purpose of the 2875-MHz profiler was to observe the vertical structure of hydrometeors in and
below the precipitating clouds that advected over the
field site. Vertical structures resolved by the profiler
include the radar bright band in stratiform rain. In ad-
dition the profiler is used for estimating the vertical air
motion, and retrieving the raindrop size distribution
(DSD). Since the profiler has a fixed dynamic range of
about 60 dBZ, the profiler operated in three different
modes to adjust the minimum detectable signal to re-
solve different precipitation features. In all three
modes, the radar beam was directed vertically to ob-
serve directly over the profiler.

The three modes of the 2875-MHz profiler were
categorized by the radar-transmitted pulse length and
by the attenuation in the radar receiver circuitry. The cloud mode used a long pulse length of 105 m and pulse
coding to increase the radar sensitivity to low-reflec-

table 2. Operating parameters for the 2875-MHz profiler. The top five rows list mode-independent parameters. The bottom rows specify mode-dependent parameters.

<table>
<thead>
<tr>
<th>Parameters</th>
<th>Cloud mode</th>
<th>Precipitation mode</th>
<th>Attenuated mode</th>
</tr>
</thead>
<tbody>
<tr>
<td>Frequency</td>
<td>2875 MHz</td>
<td>2875 MHz</td>
<td>2875 MHz</td>
</tr>
<tr>
<td>Wavelength</td>
<td>10.4 cm</td>
<td>10.4 cm</td>
<td>10.4 cm</td>
</tr>
<tr>
<td>Peak power</td>
<td>380 W</td>
<td>380 W</td>
<td>380 W</td>
</tr>
<tr>
<td>Antenna</td>
<td>1.2-m shrouded dish</td>
<td>1.2-m shrouded dish</td>
<td>1.2-m shrouded dish</td>
</tr>
<tr>
<td>Beamwidth</td>
<td>2.5°</td>
<td>2.5°</td>
<td>2.5°</td>
</tr>
<tr>
<td>Height resolution (m)</td>
<td>105</td>
<td>60</td>
<td>60</td>
</tr>
<tr>
<td>Receiver attenuation (dB)</td>
<td>0</td>
<td>0</td>
<td>28</td>
</tr>
<tr>
<td>Min sensitivity at 2 and 10 km (dBZ)</td>
<td>−24 and −10</td>
<td>−10 and 3</td>
<td>22 and 36</td>
</tr>
<tr>
<td>Dynamic range (dBZ)</td>
<td>60</td>
<td>60</td>
<td>60</td>
</tr>
<tr>
<td>Max height sampled (km)</td>
<td>15.1</td>
<td>15.1</td>
<td>15.1</td>
</tr>
<tr>
<td>Max radial velocity (m s⁻¹)</td>
<td>15.8</td>
<td>16.9</td>
<td>16.9</td>
</tr>
<tr>
<td>Pulse coding (bits)</td>
<td>8</td>
<td>0</td>
<td>0</td>
</tr>
<tr>
<td>Interpulse period (μs)</td>
<td>110</td>
<td>110</td>
<td>110</td>
</tr>
<tr>
<td>Coherent integrations</td>
<td>15</td>
<td>14</td>
<td>14</td>
</tr>
<tr>
<td>Spectra averages</td>
<td>10</td>
<td>10</td>
<td>10</td>
</tr>
<tr>
<td>Spectral points</td>
<td>256</td>
<td>256</td>
<td>256</td>
</tr>
<tr>
<td>No. of range gates</td>
<td>144</td>
<td>250</td>
<td>250</td>
</tr>
<tr>
<td>Dwell time (s)</td>
<td>15</td>
<td>15</td>
<td>15</td>
</tr>
<tr>
<td>Mode sequence*</td>
<td>1</td>
<td>2</td>
<td>3</td>
</tr>
<tr>
<td>Recording</td>
<td>Full spectra</td>
<td>Full spectra</td>
<td>Full spectra</td>
</tr>
</tbody>
</table>

* The radar continuously changes modes in the order specified in the mode sequence.
activity clouds. The *precipitation mode* used a shorter pulse length of 60 m to increase the vertical resolution. The *attenuated mode* used the same operating parameters as the precipitation mode with an additional attenuator to resolve the low-altitude high-reflectivity rain events that saturate the cloud and precipitation modes (White et al. 2000). The three modes had minimum detectable signals of $-10$, $3$, and $36$ dBZ at 10 km (cloud, precipitation, and attenuation modes). The first few rows of Table 2 describe the mode-independent operating parameters and the bottom rows describe the mode-dependent parameters.

As an example of the sensitivity of each mode, Fig. 3 shows the time–altitude cross section of reflectivity for each mode during the first observed rain event on 30–31 July. Figure 3a shows the cloud mode, Fig. 3b shows the precipitation mode, and Fig. 3c shows the attenuated mode. The cloud mode has increased sensitivity to the low-reflectivity ice clouds above 8 km. As seen in Fig. 3b, the precipitation mode has better resolution of the precipitation below and through the radar bright band located around 4.8 km. The attenuated mode has nonzero values only in the lowest range gates and in the radar bright band. Comparison of the observations in the three modes reveals that the cloud and precipitation modes were saturating at low range gates during the initial portion of the storm near 0040 UTC.

In this study, the cloud and precipitation modes are analyzed independently without any attempt for correcting for the attenuation of large reflectivities near the surface. This is not a serious issue when analyzing stratiform rain that has low to moderate reflectivity val-
ues (10–35 dBZ). A merged dataset could be constructed to account for the attenuation during high-reflectivity events at low altitudes by merging the three data modes collected over a 1-min interval (White et al. 2003). The 2835-MHz profiler reflectivity is absolutely calibrated by comparing the profiler observations with simultaneous NCAR S-Pol observations over the profiler site. The calibration procedure is described in the appendix.

b. Operating modes: 449-MHz profiling radar

The main purpose of the 449-MHz profiler for the NAME 2004 campaign was to observe the vertical air motion during precipitation events. Since it is raining less than 100% of the time over the profiler site, the 449-MHz profiler was also configured to measure the horizontal wind using the Doppler beam swinging (DBS) method. The vertical mode used a vertically directed radar beam and the DBS mode used three radar beams pointed in the north, vertical, and east directions. Both modes had vertical resolution of 250 m and temporal resolution of 40 s.

The duty cycle for the vertical mode was much greater than for the DBS mode so that the vertical air motion could be sampled more frequently during rain. The 449-MHz profiler collected five vertical mode profiles then collected one DBS triad consisting of vertical, north, and west profiles. The first five rows of Table 3 list the mode-independent 449-MHz profiler parameters and the bottom rows list parameters for the vertical and DBS modes. The 449-MHz profiler reflectivity is calibrated to the 2835-MHz profiler by comparing simultaneous reflectivity spectral densities observed in the vertical beams. Since both radars are sensitive to Rayleigh scattering from the hydrometeors, the gain of the 449-MHz profiler is adjusted until the Rayleigh scattering portions of the spectra agree between the two radars.

3. Observed vertical structure of precipitation during NAME 2004

The first objective of this project was to observe the vertical structure of precipitating cloud systems that compose the monsoon system to help understand the microphysical processes of precipitation in the monsoon. During the 2-month campaign, 23 rain events passed over the NOAA profiler site. Each rain event consisted of one or more precipitation cells that had reflectivity of at least 20 dBZ at 2 km. One example of a precipitation event is shown using the cloud mode in Fig. 4 and consisted of two closely spaced cells. The vertical structure of the precipitation is characterized using the moments of the Doppler velocity spectra, which include the total reflectivity, mean Doppler velocity, and the velocity variance. The first cell occurred near 0000 UTC on 18 August and consisted of a convective element embedded between stratiform elements as indicated by weak, yet identifiable, bright bands. The spectral variance in the first cell is larger...
than in the second cell indicating that the first cell has more turbulence. This cell had upward velocities greater than 2 m s\(^{-1}\) below 4 km that preceded the large reflectivities at this same altitude. The increased reflectivity is due to larger raindrops falling out of the updraft while the smaller raindrops are lifted in the updraft (Atlas and Williams 2003; Atlas et al. 2004). When comparing the spectrum variance between the two cells, the larger spectrum variance in the first cell indicates that this cell had more turbulent motions associated with the convective element and the transition into the stratiform elements.

The second rain cell between 0100 and 0230 UTC consisted of stratiform rain as identified by three features: the radar bright band near the freezing level in the reflectivity field, the increase in mean Doppler velocity near the freezing level as the snow and ice particles falling at 1–2 m s\(^{-1}\) melt and fall as raindrops at 6–8 m s\(^{-1}\), and by the small spectrum variance above the freezing level indicating nonturbulent air motions. Increased spectrum variance occurs after 0230 UTC above 8 km associated with possible mammatus clouds. These signatures in the vertical profile of radar moments are used to classify precipitation into different rain regimes including stratiform and convective rain (Williams et al. 1995; White et al. 2002). Each profile is classified into stratiform and nonstratiform rain using the methodology described in section 5 with the profiles classified as stratiform rain indicated with asterisks in Fig. 4. During the NAME 2004 deployment, 1401 pre-
cipation mode profiles were observed with reflectivities greater than 20 dBZ at 2 km. Of these 1401 profiles, 937 profiles (67%) had well-defined bright bands and were classified as stratiform rain while the remaining profiles were classified as nonstratiform rain.

4. Vertical air motions and raindrop size distribution profiles

The second objective of this project was to estimate the vertical profile of vertical air motions and DSDs from near the surface to just below the melting layer. The key components needed to make these retrievals are two profilers that are sensitive to two different scattering processes in the atmosphere. The 449-MHz profiler is sensitive to inhomogeneities in the refractive index caused by turbulence and the 2875-MHz profiler is sensitive to the backscattered energy from the raindrops. Combining these observations will yield profiles of vertical air motions and DSDs.

a. Estimating the vertical profile of vertical air motion

Profilers observe and record the Doppler velocity spectra at each range gate. When calibrated to the Rayleigh scattering from liquid raindrops, the Doppler velocity spectra are expressed in reflectivity spectral density units [(mm$^6$ m$^{-3}$)(m s$^{-1}$)$^{-1}$]. The total reflectivity (mm$^6$ m$^{-3}$) is determined by integrating the reflectivity spectral density over the valid velocity range (m s$^{-1}$). Figure 5 shows the observed Doppler velocity reflectivity spectral density at each range gate from one profile of the 449-MHz profiler (panel a) and the 2875-MHz profiler (panel b). To show the dynamic range of the reflectivity spectral density, the color pixels in Fig. 5 have the units 10log10[(mm$^6$ m$^{-3}$)(m s$^{-1}$)$^{-1}$]. The 449-MHz profiler spectra contain information on the atmospheric motions and the motions of the hydrometeors. Below the altitude of 4 km, there are two distinct peaks in the spectra. One peak is near zero velocity and the other peak occurs at downward velocities between 4 and 6 m s$^{-1}$. While the raindrops move upward and downward with the ambient air motion, the raindrops will be associated with the more downward spectral peak because of their downward terminal velocities. Thus, the peak near-zero velocity is associated with the ambient air motion and the peak between 4 and 6 m s$^{-1}$ is associated with the raindrops within the radar pulse volume during the radar dwell time. An algorithm to estimate the raindrop size distribution is described below and determines the best Gaussian-shaped function to fit the air motion spectral peak at each range gate. The mean air motion and the spectral width of the turbulence as quantified by the fitted Gaussian function standard deviation (STD; following Gossard 1994) are plotted in both spectral panels in Fig. 5.

In contrast to the 449-MHz profiler, the 2875-MHz profiler spectra shown in Fig. 5b do not resolve the ambient air motion, but only observe the Doppler motion of the hydrometeors within the radar pulse volume and dwell time. While both radars are sensitive to Rayleigh scattering due to the distribution of hydrometeors within the radar pulse volume, the 449-MHz profiler is also sensitive to Bragg scattering and can be used to estimate the ambient vertical air motion. Although either profiler could be used to estimate the DSD from the observed Rayleigh scattering, the calibrated 2875-MHz profiler spectra yield better DSD estimates because of the smaller radar pulse volume, higher spectra velocity resolution, higher sensitivity to smaller raindrops, and shorter dwell times when compared to the 449-MHz profiler. The lines labeled “1mm,” “3mm,” and “6mm” drawn in Fig. 5 represent the air-density-corrected terminal velocities for spherical drops having these diameters. These lines provide a visual aid in interpreting the observed spectra.

b. Estimating the vertical profile of DSDs

The vertical profiles of DSDs can be estimated from the Doppler velocity spectra profiles shown in Fig. 5. The DSD at each range gate is estimated by converting the radar-estimated reflectivity spectral density $S(\nu)$ into a raindrop number concentration $N(D)$ using the relationship

$$S(\nu)d\nu = N(D)dDdD,$$

where $\nu$ (m s$^{-1}$) and $d\nu$ (m s$^{-1}$) are the velocity and velocity resolution of the spectral density, $D$ (mm) is the mean raindrop diameter that corresponds to the velocity $\nu$, and $dD$ (mm) is the diameter range corresponding to each velocity range $d\nu$. In radar applications, the velocity resolution $d\nu$ is constant across the velocity range while $dD$ is variable and dependent on the size of the raindrop. Through laboratory studies, the transformation from velocity to diameter space in (1) is facilitated with the raindrop diameter to terminal fall speed relationship expressed as

$$V_{\text{fall speed}}(D) = [9.65 - 10.3 \exp(-0.6D)]\left(\frac{\rho}{\rho_0}\right)^{-0.4},$$

where $\rho_0$ and $\rho$ represent the air densities at the ground and the level of the observation aloft, respectively (Gunn and Kinzer 1949; Atlas et al. 1973).
Fig. 5. Reflectivity Doppler velocity spectral density observed on 31 Jul 2004 by (a) the 449-MHz profiler (0108:08 UTC) and (b) the 2875-MHz profiler (0108:34 UTC).
The Doppler velocity reflectivity spectral density observed by a radar is not just the simple expression of (1), but is the reflectivity spectral density shifted by the ambient air motion and broadened by the turbulent motions and horizontal wind motions within the radar pulse volume. For the 2875-MHz profiling radar, the observed reflectivity spectral density can be expressed as

\[ S_{\text{observed}}(v) = S_{\text{air}}(v) S_{\text{hydrometeor}}(v) + P_{\text{noise}} \]

where \( S_{\text{air}}(v) \) is the normalized Gaussian function representing the ambient air motion, \( S_{\text{hydrometeor}}(v) \) is the unknown hydrometeor reflectivity spectral density within the radar pulse volume, and \( P_{\text{noise}} \) is the noise level for the observed spectrum (Wakasugi et al. 1986).

Expanding the expression on the right-hand side of (3), the observed reflectivity spectral density can be modeled as

\[
S_{\text{model}}(v) = \frac{1}{2\pi \sqrt{\sigma_{\text{air}}}} \exp \left\{ \frac{-(v - v_{\text{air}})^2}{2\sigma_{\text{air}}^2} \right\} \times \left[ N_{0}D^{6+\mu} \exp \left(-4 \frac{D}{D_m}\right) \frac{dD}{dv} \right] + P_{\text{noise}},
\]

where \( v_{\text{air}} \) and \( \sigma_{\text{air}} \) are the estimated parameters of the Gaussian air motion function, and the DSD is described using a modified gamma function defined by the scale parameter \( N_{0} \), the shape parameter \( \mu \), and the mass-weighted mean drop diameter \( D_m \) (Ulbrich 1983). Rajopadhyaya et al. (1999), Cifelli et al. (2000), and Schafer et al. (2002) have used two collocated profilers to estimate the DSD by first estimating the Gaussian air motion function using one profiler and then estimating the three DSD parameters using the observations from the other profiler. This study uses a similar methodology by estimating the air motions using the 449-MHz profiler observations and estimating the DSD using the 2875-MHz profiler observations. An example of estimating the air motion using the 449-MHz profiler spectra is shown in Fig. 5. The air motion Gaussian function is interpolated in time and in height to match the resolution of the 2875-MHz profiler observations. The best DSD is determined by systematically varying the shape parameter and mean drop diameter over the ranges of \( 0 \leq \mu \leq 20 \) and \( 0.25 \leq D_m \leq 3.5 \) mm and adjusting the scale parameter for each \( (\mu, D_m) \) pair until the difference between the modeled spectrum and the observed spectrum is minimized using a least square fit criterion.

The retrieved profile of DSD parameters can be used to estimate profiles of precipitation liquid water content or total rain rate as a function of altitude. These parameters can also be used to construct distributions of liquid water content \( M(D) \) or rain rate \( R(D) \) as functions of raindrop diameter. These distributions can be expressed as

\[
M(D) = N_0D^{3+\mu} \exp \left(-4 \frac{D}{D_m}\right)(\text{g m}^{-3})(\text{mm}^{-1})
\]

and

\[
R(D) = V_{\text{fall speed}}(D)N_0D^{3+\mu} \exp \left(-4 \frac{D}{D_m}\right)(\text{mm}^{-1}) \times (\text{mm}^{-1}),
\]

and the precipitation liquid water content (LWC) and total rain rate (RR) are simply the integrals of (5) and (6) over all raindrop diameter sizes. The liquid water content densities for the 2875-MHz profiler observations shown in Fig. 5 were estimated for each range gate below 3.5 km and are plotted in Fig. 6. The DSD estimation is performed at each range gate without any input from retrievals performed at other range gates. Thus, continuity in altitude is a result of the continuity of the DSD with altitude and not a consequence of the retrieval method. Note that the raindrop size distributions are narrower above the altitude of 2 km and broader below this altitude. The increase in drop size indicates some drops are coalescing and some drops are experiencing breakup to maintain the number of small drop sizes. The discontinuity at 0.75 km is hard to diagnose with a single profiler and is probably due to advection of a different rain cell over the profiler.

5. Extending the view beyond NAME tier I using TRMM observations

While the ground-based profiling radar continuously looks upward and observes the precipitating cloud systems that pass over the profiler site, the Tropical Rainfall Measuring Mission (TRMM) Precipitation Radar (PR) continuously looks downward and observes precipitating cloud systems that the satellite passes over. The TRMM PR enables the precipitation vertical structure to be analyzed beyond the NAME tier I region and to address the third objective of this project, which is to better understand the microphysical processes of oceanic and continental rain regimes.

The TRMM satellite completes about 17 earth orbits during each 24-h period and will resample the same location every couple of days. There is a very low probability that the PR will pass over the profiler site while both radars are observing precipitation. During the
2-month NAME 2004 campaign, the PR passed over the NOAA profiler site 24 times, and there were no overpasses with simultaneous rain observations.

The closest we came to having a simultaneous profiler–TRMM rain observation occurred on 18 August. The profiler observations for this event are shown in Fig. 4. The TRMM PR passed over the NOAA profiler site at 0305 UTC and the PR reflectivity at 2 km is shown in Fig. 7a. There would have been simultaneous profiler and PR observations of precipitation if the overpass had occurred 1 h earlier. The measured profiles of 13.8-GHz PR reflectivity with 250-m-range resolution are corrected for precipitation attenuation (Iguchi et al. 2000) and are available through the National Aeronautics and Space Administration (NASA) Physical Oceanography Distributed Active Archive Center (PO.DAAC) in the version 6 2A25 data product. The PR observations are classified by NASA into convective and stratiform rain regimes based on the vertical structure and the spatial distribution of the reflectivity and are available in the version 6 2A23 data product. A description of these algorithms is available online from the NASA PO.DAAC (TRMM Precipitation Radar Team 2005). Figure 7b shows the convective and stratiform classifications for each rain pixel during this PR overpass. Notice that the convective pixels tend to have larger reflectivities and large spatial reflectivity gradients compared with the stratiform regions. These two
signatures are used in the spatial distribution classification algorithm to identify convective rain.

The cross section labeled A–B drawn in Figs. 7a and 7b is drawn through stratiform and convective regions of the storm and through the profiler site, which is shown with a red symbol. The vertical profiles of PR reflectivity along the A–B cross section are shown in Fig. 7c. The convective–stratiform classification from the version 6 2A23 algorithm is shown along the top portion of Fig. 7c. On both sides of the profiler site, the PR classifies the precipitation as stratiform rain. On the A side of the profiler, the PR reflectivity profile has a well-defined bright band. But on the B side of the profiler, the PR reflectivity profile does not show a bright band, yet these profiles are classified as stratiform rain. As discussed in detail below, this stratiform classification results from the precipitation extending above the freezing level and not having large spatial gradients. Thus, these profiles are stratiform rain even though a bright band is not observed.

The extremely low probability of simultaneous surface profiling radar and satellite PR observations of precipitation prevented a direct comparison of reflectivities and rain regime classification methodologies between the two instruments during NAME 2004. But statistical comparisons can be made by comparing the reflectivity distributions made at a single field site with the reflectivity distributions from the satellite observations around that field site. In conducting statistical comparisons, the question implicitly being addressed is whether the statistics of both instruments could be drawn from the same sample population.

While the vertical structure of precipitation is different for stratiform and convective rain due to the microphysics of these two rain regimes (Houze 1993), the climatological spatial distribution of these two regimes
within the NAME 2004 domain is not actually known. To quantify the regional dependency of the spatial distribution of precipitation and to analyze year-to-year variability, the domain around the NOAA profiler site was divided into 1° × 1° grid boxes. Three regions are shown in Fig. 8 and are defined as gulf, coast, and land regions. The gulf region consists of grid boxes predominantly over the Gulf of California. Each coast region grid box contains both gulf and land surfaces. And the land region contains grid boxes that are over land and includes very few mountainous regions.

Since the vertical structure of reflectivity is a direct indication of the microphysical processes occurring within that column of rain, every profile of reflectivity retrieved by the TRMM Precipitation Radar (PR) is classified as either stratiform or convective rain. Every profile of reflectivity is further analyzed to determine if it should also be a member of the bright band exists or the shallow rain partitions (or subclassifications). While the actual classification of the PR observations is a thorough process performed by the PO.DAAC, it is simply described here as a four-step process. First, a search for a radar bright band is performed on every profile of reflectivity. If a bright band does exist, then that profile becomes a member of the bright band exists partition and is also classified as stratiform rain. The texture of the reflectivity across the spatial extent of the rain field is analyzed to identify regions of large reflectivity and regions with large reflectivity gradients. The profiles of reflectivity in these regions are identified as convective rain. The remaining profiles of reflectivity that have not yet been classified are assigned to the stratiform rain classification. The fourth step analyzes the vertical extent of the profile of reflectivity. If the profile of reflectivity is much lower than the estimated height of the freezing level (TRMM Precipitation Radar Team 2005), then that profile becomes a member of the shallow rain partition. From this methodology, shallow rain profiles can have either stratiform or convective rain classification and bright band exist profiles are a subset of the profiles classified as stratiform rain.

The ground-based vertically pointing profiling radar observations can be classified using some, but not all, of the methods used to classify the PR observations. Although the ground-based system cannot use the instantaneous horizontal gradient information to identify the convective rain regions, the profiling radar can use the brightband identification algorithm to identify the bright band exists partitioned profiles. To this end, the profiler observations that contain precipitation (Ralph et al. 1995, 1996; Williams et al. 2000) were analyzed to identify all reflectivity profiles with well-defined bright bands, and the resulting reflectivity distribution at each altitude is shown in Fig. 9b. Since the brightband altitude changes as the freezing level changes, the altitudes of each range gate are plotted relative to the altitude of the maximum reflectivity observed in the bright band. This normalization ensures that the altitudes of the reflectivity distributions are relative to a common physical phenomenon. The colors shown in Fig. 9b represent the values of the probability distribution function (PDF) of reflectivities at each altitude with the total number of observations at that altitude shown in Fig. 9a.

The reflectivity distributions constructed from PR profiles classified as containing a bright band are shown in Figs. 9c, 9d, and 9e for each region during the 2004 season. The number of land region PR observations decreases below the brightband reference altitude because the freezing level is closer to the surface in the mountainous areas. It can be seen in Fig. 9 that the vertical structure of the reflectivity distributions is very similar for the ground-based radar and satellite-based radar across the three regions in the NAME domain.

Figures 10a and 10b show the details of the reflectivity distributions at +2 and −2 km from the brightband reference altitude, respectively. The PR reflectivity PDFs for each region are shown using symbols and thin lines while the profiler reflectivity PDF is shown using the thick line. Above the bright band, the profiler PDF of reflectivity agrees very well with the PR PDFs of reflectivity for the three different regions. Below the
bright band, the profiler PDF has the same mode as the PR PDFs, but the tails of the profiler distribution are skewed to smaller values because the profiler has greater sensitivity than the PR. While possible reasons for the differences are discussed in section 6, the general agreement in PDF statistics implies that the 2-month-long ground-based observations captured the features observed by the satellite around the ground-based station during NAME 2004.

6. Climatological context using TRMM PR

In the previous section, the TRMM PR reflectivity distributions and the profiler reflectivity distributions were compared for July and August of 2004. In this section, the PR observations in the gulf, coast, and land regions are analyzed for eight July–August seasons from 1998 through 2005 and the resulting bulk statistics are given in Fig. 11. Figure 11a shows the observation number density (defined as the total number of PR observations within each region normalized by its area) for the gulf, coast, and land regions. More rain events occurred over the land region than over the coastal and gulf regions, which is consistent with the more frequent rain events occurring at the highest terrain of the Sierra Madre Occidental as observed by the NAME Event Rain Gauge Network (NERN; Gochis et al. 2004). Although the total number of 2001 observations is underestimated due to the lack of measurements for 16 days as the satellite performed a repositioning maneuver to change the nominal orbit altitude from 350 to 402 km, there is some interannual variation in the number of rain observations over this 8-yr period. Also, there is not a step change in the pre-2001 and post-2001 statistics, indicating that the boost in the TRMM satellite altitude does not affect these statistics.

The gulf region had the fewest and the land region had the most rain observations. Conversely, all three...
regions had similar frequencies of occurrence of stratiform and convective rain. Figure 11b shows the stratiform rain frequency of occurrence for each region and for each year as well as the frequency of occurrence of the reflectivity profiles with well-defined radar bright bands. Figure 11b shows that the gulf region tended to have slightly higher frequencies of occurrence of stratiform rain and profiles with a bright band than the other regions. Figure 11b also shows that the year-to-year variability of these percentages within each region is greater than the differences between the three regions. Thus, while Fig. 11a shows that the rain event number density is highest over the land region, Fig. 11b shows that the frequencies of occurrence of stratiform rain and reflectivity profiles with well-defined bright bands are similar for the three regions.

Figure 12 shows the reflectivity probability distribution functions for profiles with well-defined bright bands using the PR observations accumulated over the 1998–2005 seasons similar to those shown in Fig. 9 for the 2004 season. The altitude of the individual profiles has been normalized relative to the altitude of the bright band. There are several similarities between the profiler and multiyear PR profile of PDFs. First, all four distributions have narrow distributions of reflectivities above the bright band that result from a narrow size distribution of falling ice and snow particles. Also, all four distributions show a well-defined bright band. Below the bright band, the reflectivity distribution mode decreases slightly, presumably due to evaporation, as the rain falls through drier air.

Figures 13a and 13b show the details of the reflectivity distributions at +2 and −2 km away from the bright band, respectively. The 8-yr mean reflectivity PDF for each region is shown by symbols and the year-to-year variability is shown by vertical bars representing one standard deviation of the annual PDF values. The symbols are offset by 0.15 dB in the plot to help visualize the STD lines. Note that the PDFs for the different regions are similar with the mean values within the year-to-year variability of the three regions. Thus, the year-to-year variability in reflectivity within each region is larger than the variability between regions.

Also shown in Figs. 13a and 13b are the PDFs of reflectivities observed by the ground-based profiling radar during the NAME 2004 campaign. The general features of the multiyear dataset are similar to the features shown in Fig. 10 for the 2004 season. Specifically, above the bright band, the PDF of the profiler agrees very well with the PR PDFs for the three different regions. Below the bright band, the profiler PDF has the same mode as the PR PDFs, but the tails of the profiler distribution are skewed to smaller values. The disagree-
ment for reflectivities less than 20 dBZ is probably due to the different minimum sensitivities of the two instruments with the PR limited to reflectivities greater than about 15 dBZ. The disagreement for reflectivities greater than 30 dBZ may be due to the bright band interfering with the PR attenuation correction, which may cause the reflectivity to be overestimated. Also, any misclassification of PR convective rain profiles as stratiform rain profiles would occur at larger reflectivities and cause the PR distributions to have more frequent high-reflectivity observations. While more work is needed to quantify the differences in these reflectivity PDFs below the brightband altitude, the general agreement in the PDF statistics implies that the 2-month-long ground-based observations captured the features observed by the satellite during 8 yr of observations around the ground-based station.

7. Concluding remarks

The NAME 2004 field campaign was held from June to September 2004 with observations in northwest Mexico designed to observe the environment around and within the North American monsoon. The NOAA/Aeronomy Laboratory and the NOAA/Environmental Technology Laboratory maintained a field site about 100 km north of Mazatlán. This NOAA profiler site used wind profilers, precipitation profilers, ground-based upward- and downward-looking radiometers, and a 10-m meteorological tower to measure the attributes of the monsoon circulation and embedded clouds and precipitation.

The precipitation profiler operated at 2875 MHz and observed the vertical structure of the precipitation. When combined with the vertical air motions estimated by the 449-MHz profiler, the raindrop size distribution can be estimated from near the surface to just below the freezing level. The absolute calibration of the precipitation profiler was achieved by using the NCAR S-Pol scanning radar observations over the NOAA profiler site as described in the appendix.

To place the single-season observations into context with other monsoon seasons, the precipitation profiler reflectivity distributions were compared with 8 yr of
TRMM Precipitation Radar (PR) reflectivity distributions. Since there were no simultaneous surface-based and satellite-based precipitation radar observations during NAME 2004, a statistical analysis was used to evaluate whether the reflectivity distribution observed at the NOAA field site is similar to the reflectivity distributions from regions around the field site and for eight consecutive seasons.

The statistical analysis revealed that the number of rain events observed by the TRMM PR is dependent on geography since the PR observed more events over land than over the coast and gulf regions. But interestingly, the frequencies of occurrence of stratiform rain and of reflectivity profiles with radar bright bands are essentially independent of region. Also, the distribution of the reflectivity at each height has more year-to-year variability than region-to-region variability. These findings lead to the conclusion that the microphysical processes associated with rain with a well-defined bright band have a larger influence on the vertical structure of reflectivity than the regional location of that rain profile. In other words, once a well-defined bright band has developed, the vertical profile of reflectivity and its distribution at individual altitudes is similar over the gulf, coast, and land regions. This analysis does not address the generation or the life cycle of profiles displaying bright bands. Further work is needed to place the brightband profiles into a precipitation-regime-dependent context or into a precipitation life cycle context, which are dependent on regional and large-scale dynamical forces.

Acknowledgments. This work was supported in part by NOAA’s Office of Global Programs under the GEWEX Americas Prediction Project (GAPP) through Grant GC04-064 and supported in part by NASA’s Precipitation Measurement Mission. Appreciation is extended to Scott Abbott, Dave Costa, Dan Gottas, Jim Jordan, Clark King, Jesse Leach, David White, and Robert Zamora for installing and maintaining the hardware and collecting the observations during the NAME 2004 campaign. Thank you to Linda Benson for editing the manuscript.

APPENDIX

Calibration of Profiler Using S-Pol

In past experiments, a surface disdrometer has been used to absolutely calibrate the profiling radar (Williams et al. 2005; Gage et al. 2000, 2004). But since the Joss–Waldvogel disdrometer (Disdromet, Inc.) deployed for NAME 2004 did not function properly, the well-calibrated NCAR S-Pol scanning radar located 45 km away was used to absolutely calibrate the profiling radar. One of the difficulties in comparing scanning radar and profiling radar observations is that the two instruments have different temporal and sampling resolutions. To address the different temporal resolutions, the scanning radar made special range–height indicator (RHI) observations over the profiler site when the on-duty S-Pol scientist thought that there was rain over the profiler site. These special RHI scans occurred in between the routine volume scans. Thus, the temporal resolution of the scanning radar observations over the profiler site consisted of nearly instantaneous observations acquired every few minutes. This is in contrast to the profiler temporal resolution of 15-s dwells repeated for each mode every 45 s. More details of the S-Pol radar operating during the NAME 2004 campaign can be found in Lang et al. (2007).

The 2875-MHz profiler and S-Pol reflectivities over the NOAA profiler site are shown in Figure A1 for the 30–31 July rain event. Figure A1a shows the profiler precipitation mode reflectivity at the original 60-m vertical resolution and 15-s dwell time. Figure A1b shows the profiler reflectivity after degrading the vertical resolution to 780 m to match the S-Pol vertical resolution over the profiler site due to the 1° S-Pol beamwidth. Figure A1c shows the S-Pol reflectivity directly
over the profiler site collected from the RHI scans and the volume scans. The shift in time with each S-Pol observation results from the time delay between each elevation scan as the S-Pol radar completes each 15-min volume scan. Even though the S-Pol observations are nearly instantaneous over the profiler site, the observations are shown with a 5-min dwell to enable the reflectivity pseudocolor to be plotted in the panel. The increased number of S-Pol observations near 0100 UTC are from the special RHI scans requested in real time by the S-Pol scientist.

To calibrate the profiler reflectivity with the S-Pol observations, the closest 15-s dwell time and 780-m vertically averaged profiler observation were compared with spatially averaged S-Pol observations over the profiler site. Each 780-m vertically averaged profiler observation was constructed by linearly averaging the profiler reflectivity at the thirteen 60-m range gates centered at the middle of the S-Pol beamwidth. The S-Pol reflectivities for each elevation scan are averaged over a 750-m-diameter circle centered over the nominal profiler location to account for misalignment and range gate to range gate variability. The 750-m diameter corresponds to three S-Pol range gates. There is still a mismatch in the pulse volumes between the two radars due to the geometry of the radar beams. This mismatch would not be a problem if the precipitation were uniform in time and space. The spatiotemporal inhomogeneity of the precipitation contributes to the representativeness of each instrument’s estimate of the precipi-

---

**Fig. A1.** Simultaneous profiler and NCAR S-Pol scanning radar observations over the NOAA profiler site. (a) Profiler observations at their original vertical resolution, (b) profiler observations reduced to match the S-Pol 780-m vertical resolution, and (c) the S-Pol observations over the NOAA profiler site.
tation and will contribute to the uncertainties in the calibration (Williams et al. 2005).

Figure A2 shows the calibration scatterplot for the 2875-MHz profiler precipitation mode using the S-Pol radar as the reference. The observations above 1.0 km were used in Fig. A2 to avoid saturation of the profiler observations. Only observations during stratiform precipitation with a well-defined bright band are used to calibrate the profiler. The gain of the 2875-MHz profiler was adjusted so that the mean difference between the profiling and scanning radar observations during stratiform rain was zero. The standard deviation of the difference in reflectivity was 2.46 dBZ and represents the uncertainty resulting from the representativeness of the instruments observing different portions of the rain due to their different spatial sampling and temporal resolution. Work is still being conducted to determine how the representativeness between a scanning radar and a profiling radar located 45 km away can be scaled to larger areas that are used in hydrologic and climate models.

REFERENCES


