ARMAR OBSERVATIONS OF THE MELTING LAYER DURING TOGA COARE

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Abstract

The NASA/JPL Rain MApping Radar (ARMAR) was operated on the NASA DC-8 aircraft during all 13 flights for TOGA COARE, acquiring data in tropical mesoscale convective systems. Processing and calibration of the ARMAR data are discussed, followed by a presentation of ARMAR measurements of the melting layer in the stratiform rain region of mesoscale convective systems.

1 Introduction

The Airborne Rain MApping Radar (ARMAR) was developed by NASA and the Jet Propulsion Laboratory (JPL) to support the Tropical Rain Measuring Mission (TRMM) [1] and also to serve as a testbed for future spaceborne rain radars. ARMAR operates on the NASA DC-8 aircraft, which flew 13 missions in January and February 1993, during the Tropical Oceans Global Atmosphere Coupled Ocean Atmosphere Response Experiment (TOGA COARE) [2]. ARMAR acquired data in a variety of tropical mesoscale convective systems (MCSs), including a tropical cyclone. This paper briefly reviews the operation of ARMAR and discusses processing and calibration of the TOGA COARE dataset. ARMAR reflectivity, Doppler, and dual-polarization measurements during TOGA COARE are then used to characterize the structure and microphysics of the melting layer in the stratiform, or anvil, precipitation region of tropical MCSs. This region is produced by the sinking and melting of ice particles in regions of low vertical air velocity [3]. While the rainfall rate in the stratiform region is typically much lower than in the convective, or high vertical velocity, region, stratiform rain often accounts for 50% or more of the total rain associated with MCSs because of its large areal extent [3,4]. The stratiform region is also of interest because of its affect on dynamics through cooling associated with melting [5,6,7] and because of its affect on electromagnetic waves for radars [8] and for satellite communications [9]. Observations of the reflectivity, polarization parameters, vertical velocity, and cooling in the melting layer are presented.

2 ARMAR TOGA COARE Dataset Description

ARMAR operates at 13.8 GHz, the frequency planned for the Precipitation Radar (PR) on TRMM. ARMAR simulates the TRMM PR geometry by looking downward and scanning its antenna $\pm 20^o$ in the cross-track direction. It can also acquire data with the antenna in a fixed position, usually nadir. In either antenna mode, the along-track antenna pointing is normally as close to nadir as possible, to minimize the bias in Doppler velocity due to the aircraft motion. ARMAR has capabilities in addition to those of the TRMM PR, including multiple polarization and Doppler velocity measurement.

While operating as a radar, a small fraction of time is spent measuring brightness temperature in a radiometer mode at the same frequency, viewing geometry, and resolution as the radar mode. ARMAR characteristics are shown in Table 1, and additional details can be found in [10]. During TOGA COARE, a majority of data were collected with ARMAR operating in single-polarization Doppler mode. However, a large data volume was also acquired in dual-polarization Doppler and single polarization non-Doppler modes. In the single-polarization non-Doppler mode, frequency diversity is used to remove correlation between return from adjacent pulses, increasing the number of independent samples. Dual-polarization Doppler operation uses a sequence of alternating polarization combinations, either HH-HV or HH-VV.

The ARMAR TOGA COARE data were processed in two stages. Because the transmitted signal is a frequency modulated chirp, the first processing stage performs pulse compression and computes averages of various second order statistics. Which statistics are computed depends on the mode in which data were acquired. In the second stage of processing, the pulse compressed and averaged data are passed through a program which cleans up, calibrates, and time stamps all data. The average total power is first converted to average signal power by subtracting an estimate of the system noise level. The result is converted to reflectivity factor Z_e using in-flight calibration loop measurements and laboratory system loss measurements. Use of the calibration loop is described in [10] and has the advantage of removing any biases caused by changing transmit power or receiver gain. For Doppler mode data, the pulse-pair algorithm [11] is used to estimate the mean velocity and spectrum width. The Doppler velocity has been corrected for aircraft motion by subtracting the measured surface velocity from all range bins. For dual-polarization modes, appropriate polarimetric quantities (e.g., HH-VV phase difference) are calculated, in addition to reflectivity and velocity.

3 TOGA-COARE Data Quality Assessment

Before deployment in TOGA COARE, ARMAR was field tested in a ground-based configuration during several occasions in 1991-1992 and was flight tested on the NASA DC-8 aircraft in May of 1992 and again in December 1992, just prior to the TOGA COARE deployment. These tests verified the ARMAR characteristics in Table 1 and are described in [10]. Just before TOGA COARE, a set of the laboratory system loss measurements used for radar calibration were made. These were repeated after TOGA COARE and were essentially identical to the first set.

Calibration was checked by examining the measured ocean backscatter cross section σ^o under clear conditions. A 10^o incidence angle was chosen because σ^o has the least sensitivity to the surface wind speed, decreasing slightly with increasing wind speed [12]. The average σ^o_{HH} over all flight days was 7.7 dB, with a standard deviation of 1.2 dB. This is similar to the measurements presented by Schroeder et al. [12]. For example, their measured σ^o_{HH} at 10^o incidence, looking downwind, is 7.9 dB at 2 m/s wind speed and 7.2 dB at 5 m/s. The accuracy of ARMAR Doppler velocity measurements was determined by comparing the measured ocean surface velocity with the surface velocity predicted from the ARMAR antenna angle and the aircraft roll, pitch, yaw, and velocity. The surface velocity is normally non-zero because the antenna pointing vector often had a small component in the flight direction. Using data from all flight days, the mean velocity difference (predicted minus measured) is only 0.01 m/s. The RMS difference between predicted and measured velocities is 2.08 m/s. Surface velocities used in the analysis ranged from 0 m/s up to 25 m/s. The precision of the ARMAR velocity measurements was estimated using the approximate analytical result for the pulse pair estimator in [11], using Monte Carlo simulation, and by fitting the surface velocity across a scan to the theoretical dependence of the Doppler velocity on antenna scan angle. All three methods indicate that for moderate to high SNR measurements, the ARMAR velocities should have a precision of 0.3 m/s.

4 Melting Layer Observations

Figures 1 and 2 show typical ARMAR multi-parameter profiles through the melting layer in stratiform precipitation during TOGA COARE. The radar equivalent reflectivity Z_e near the 0^o C isotherm sharply increases with decreasing altitude as ice particles begin to melt. The maximum reflectivity occurs at around 4.5 km in both cases. The region of enhanced reflectivity associated with the melting layer is called the bright band and was first noted many years ago (see, e.g., [13]). The velocity of hydrometeors v above the bright band is downward at 1-2 m/s and increases to 7-8 m/s below the bright band. The data in Figure 1 were acquired with HH-HV dual-polarization, and the Linear Depolarization Ratio (LDR), defined as the HV/HH backscatter ratio, is plotted along with Z_e and v. LDR reaches its maximum value slightly below the maximum Z_e. The data in Figure 2 were acquired using HH and VV dual-polarization, and the HH-VV correlation coefficient ρ_{HV} is plotted. It reaches its minimum value somewhat below the maximum of Z_e.

To better understand the radar characteristics of the melting layer, the statistics of ARMAR nadir-looking observations for one or more flight lines for each flight day (except Julian day 12) were examined. On the first flight, Julian day 12, no stratiform precipitation was encountered. Table 2 shows the mean and standard deviation of 10 parameters. Also shown are the number of points that went into the calculation of the mean and standard deviation. Each point corresponds to an area at the surface with diameter of 800 m, and points are separated by at least 400 m. As can be seen in Table 2, many fewer points were available for the dual-polarization parameters.

The first parameter in Table 2 is the maximum equivalent reflectivity Z_e of the bright band. It ranged during TOGA COARE from around 20 dBZ to 47 dBZ, with a mean of 36 dBZ. The altitude at which the bright band reflectivity reaches its maximum value is the second parameter in Table 2. It varied between 4.0 and 5.1 km, with a mean of 4.5 km. The DC-8 aircraft was equipped with a dropsonde system to allow profiles of atmospheric variables to be made at selected locations. The third parameter in Table 2 is the dropsonde measured altitude of the zero degree isotherm. The mean level was 4.8 km. The mean bright band altitude was approximately 300 m below the mean 0^o C isotherm.

In examining the statistics of the maximum reflectivity and the maximum reflectivity altitude, it was noted that the two parameters are correlated under certain conditions. Figure 3 is a scatter plot of the maximum bright band reflectivity Z_e versus the altitude for Julian days 37 and 51, showing that higher reflectivities typically occured at lower altitudes. This is true of the other days as well; bright band altitude and maximum reflectivity are negatively correlated at the 1% significance level. However, if the data from multiple flight days are analyzed together, the correlation is lost, probably because of synoptic scale changes in the height of the 0^o isotherm. Figure 4 illustrates the reduction in the bright band altitude with increasing reflectivity for a mesoscale convective system investigated on Day 51. Figure 4 shows the observed reflectivity versus altitude for both low and high reflectivity cases. The high reflectivity case is substantially thicker and reaches its maximum reflectivity several hundred meters below the altitude of maximum reflectivity for the low reflectivity plot. The high reflectivity data were acquired about 3 minutes after the low reflectivity data; the spatial separation is around 40 km.

Parameters 4 and 5 in Table 2 are the measured Doppler velocities of ice and rain hydrometeors. For stratiform conditions the air motion is small, so the Doppler measurements are taken to be the particle fall speeds. Parameter 4 is the fall speed of ice particles 900 m above the bright band maximum reflectivity. The mean fall speed of 1.66 m/s is typical of snow. A few measurements indicated fall speeds of greater than 3 m/s, indicating either more dense particles such as graupel or the presence of downward air motion. Parameter 5 is the maximum velocity reached during the melting process, while parameter 6 is the altitude at which the maximum was reached, relative to the altitude of maximum reflectivity. The negative value of the mean altitude indicates that the maximum velocity was reached at a lower altitude than the maximum reflectivity.

Dual-polarization measurements are shown as parameters 7-10 in Table 2. The altitude parameters are relative to the altitude of the maximum reflectivity. The negative mean values indicate that the extreme values of LDR and ρ_{HV} are usually reached at an altitude below the altitude at which Z_e reaches its maximum. This phenomenon was also noted for the cases presented in Figures 1 and 2.

5 Discussion

As noted previously, the existence of the bright band in radar observations of stratiform rain has been known for many years. The primary cause is the increase in the dielectric constant of melting hydrometeors, followed by an increase in the fall velocity as melting is completed [6],[13],[14]. Other factors noted in these references which may also have an effect include growth of the ice phase particles via deposition, aggregation of wet ice phase hydrometeors, breakup of raindrops after melting, and changes in particle size, shape, and density.

As an aid to understanding the behavior of the various radar parameters listed in Table 2, we consider scattering from a single melting snow particle. We neglect particle growth via deposition and aggregation and raindrop breakup. The melting process is parameterized by the mass fraction q of liquid water, which is zero for the snow particle at the top of the bright band and unity when the particle collapses into a raindrop. The dielectric constant for snow is found using the Maxwell-Garnet theory for an ice matrix with air inclusions. For the melting particle, the dielectric constant is found as a function of q by again using the Maxwell-Garnet theory for a water matrix with snow inclusions [15]. The equivalent diameter D of the particle as a function of q is given by [16]

$$D = D_w(q + (1 - q) * \rho_w/\rho_s)^{1/3} \tag{1}$$

where ρ is density, the subscript w denotes water, and the subscript s denotes snow. The snow particle is assumed to be a prolate spheroid with random orientation in the horizontal plane. As the particle melts, the axial ratio r decreases, reaching unity when the particle collapses into a raindrop. Laboratory observations [17], [18] indicate that the particle tends to maintain its shape as melting proceeds, collapsing into a sphere rather suddenly. This can be modeled by using

$$r = r_s + (1 - r_s)q^8 \tag{2}$$

This dependence was found from radar observations in [19]. The exponent 8 causes r to remain close to r_s until q nears unity. As q nears unity, r rapidly approaches unity, similar to the observations in [17, [18]. The particle fall velocity v is taken from a curve presented in [18], which we approximate by

$$v = v_s + (v_w - v_s) * q^3 \tag{3}$$

where v_s is the terminal velocity of the snow particle from [20] and v_w is the terminal velocity for a raindrop is given by $9.65 - 10.3\exp(-0.6D_w)$, used in [19]. Finally, we relate q to the distance y that the particle has fallen below the 0^o C level, making comparison with the ARMAR observations more straightforward. the results in [18], we take q to be approximately proportional to h, with the constant of proportionality depending on the particle size, i.e., $q = y/y_m$, where $y_m = cD_w$. The constant c is taken to be 150, with D_w in mm and y in m.

Using this simple model for a melting snow particle, the scattering matrix for the particle can be calculated as a function of q using the T-matrix volume integral equation approach [21]. This yields the scattering matrix for the particle with its symmetry axis parallel to the incident vertically polarized field. Once the canonical scattering matrix is found, the scattering matrix for an arbitrary orientation in the plane normal to the incident wavevector can be found by performing a rotation of the canonical scattering matrix. From this we can compute various second order statistics by assuming that the orientation angle is uniformly distributed and taking ensemble averages. This procedure has been used previously (see, e.g., [22]).

Figure 5 shows computations for $\rho_s = 0.1$ g/m^3 and $r_s = 3$ for particles with D_w of 2 and 5 mm. The curve HH1 is the backscatter cross section (normalized to have 0 dB maximum). Z_e, however, depends on the number density as well as the backscatter cross section. By continuity, the number density is inversely proportional to the terminal velocity v, so a quanitity that is more analogous to Z_e is the backscatter cross section divided by v. This is plotted as curve HH2. The correlation coefficient ρ_{HV} was also calculated (not shown) and reaches its minimum at the same y as the LDR maximum. The terminal velocity maximum occurs at $y = y_m$, when the particle completely collapses into a spherical raindrop.

The calculations in Fig. 5 show a behavior that is similar to the observations presented in the previous section. The maximum backscatter for the 5 mm particle is more than 10 dB greater than for the 2 mm particle, the thickness of the bright band for the 5mm particle is greater than for the 2 mm particle, and the y for which HH2 reaches its maximum is greater (i.e., lower altitude) for the 5 mm particle. This is similar to the ARMAR observations of thicker bright bands and maximum reflectivity at lower altitudes for more reflective bright bands. This phenomenon can also be seen in the observations reported in [14]. The curve HH2, which is analogous to Z_e, reaches its maximum in Figure 5 at y that is smaller (higher altitude) than the y at which the extreme values of LDR, ρ_{HV}, and v occur. This was also observed in the ARMAR data and in [23] and [24]. The reason for the upward (smaller y) displacement of the maximum of HH2 is the inclusion of the effect of velocity on the number density. Little offset is seen for HH1. Noting this, we examined the ARMAR data and found that the magnitude of the offset between the Z_e and LDR maximum or ρ_{HV} minimum is positively correlated with the terminal velocity in rain at the 1% level. Larger terminal velocities correspond to greater upward displacements of the maximum reflectivity from the maximum LDR and minimum ρ_{HV}.

In the model presented above, changes in the zero degree altitude were not considered as a contributor to the reduction in altitude with increasing maximum reflectivity. To examine this possiblity further, cases with dropsonde measurements both outside the rain area and in the rain area were examined. Out of 19 such cases, 5 were found in which the 0^o level inside the rain was at least 100 m below the level outside the rain area. The maximum depression of the 0^o level was 500 m, occurring in a case with a maximum Z_e of number of cases. This could be related to melting, since melting has been previously observed to create a nearly isothermal layer below the 0^o isotherm [6] in stratiform rain. Its cooling effect was also noted in [7]. The cooling due to melting can be estimated using the technique in [5]. The liquid water just below the bright band is assumed to have been produced entirely by melting of the ice above the melting layer. Hence, using the radar estimated liquid water content, along with the bright band thickness and rainfall velocity, the cooling rate can be estimated. The liquid water content is derived using the rainfall retrieval algorithm described in [25]. The algorithm uses a Markov model for the rain, allowing application of an extended Kalman filter. We applied the technique to a selected case for each day. Results are shown in Table 3, which gives the day, the bright band maximum reflectivity, the reflectivity of the rain just below the bright band, the liquid water content just below the bright band M, and the estimated cooling rate $\Delta T/\Delta t$. As expected, cases with larger maximum Z_e have larger cooling rates. These larger cooling rates could, over time, depress the 0^o isotherm level more significantly than the lower cooling rates associated with less reflective bright bands. It should be noted that the cooling rates in Table 3 are similar to those found in [5] and [7] and range from less than 1 K/h for light rain up to 6 K/h for the heaviest stratiform precipitation. Cases with maximum Z_e close to the mean value in Table 2 (35.4 dBZ) have cooling rates near 1 K/h.

6 Conclusions

The ARMAR airborne radar system collected a large dataset in mesoscale convective systems in TOGA COARE. The ARMAR TOGA COARE dataset was found to be of high quality and was used to study radar characteristics of the melting layer in detail over the whole of TOGA COARE. Examination of the bright band reflectivity and altitude showed that higher reflectivity is associated with lower altitude. The fall speed of ice particles just above the melting layer indicate that the particles are usually lower density ice particles, such as snow. Dual-polarization observations confirmed that particles reach maximum reflectivity and maximum LDR (or minimum ρ_{HV} at different altitudes. A simple melting layer model also produced this offset, caused by the reduction in particle number density due to the increase in fall velocity. A correlation between fall velocity and offset between the Z_e maximum and extrema of the polarization parameters was found in the ARMAR data.

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Table 1: ARMAR System Characteristics

Operating Frequency	13.8 GHz
Surface Horizontal Resolution	800 m
Vertical 6 dB Resolution	80 m
Polarization	HH,VV,HV
Antenna Beamwidth	3.8^o
Antenna Scan Angle	$\pm 20^o$
Transmitter Power	200 W
Transmit Chirp Duration	5-45 μsec
Pulse Repetition Frequency	1-8 kHz
Chirp Bandwidth	4 MHz
A/D Sampling Rate	10 MHz
A/D Resolution	12 bits
Range Sidelobes	<-55 dB
Antenna Sidelobes (2-way)	<-60 dB
Cross-pol Isolation (2-way)	-28 dB
Radiometer Δ K	1K

Table 2: TOGA COARE Melting Layer Parameters

	Parameter	Mean	Standard Deviation	Number Samples
1	Max Z_e (dBZ)	35.4	5.9	5764
2	Altitude Max Z_e (m)	4482.8	167.2	5764
3	Altitude 0^o C (m)	4884.4	182.0	73
4	Ice Particle Velocity (m/s)	1.66	1.14	5379
5	Maximum Velocity (m/s)	7.64	1.68	5379
6	Relative Altitude Max Vel (m)	-369.8	123.1	5379
7	Maximum LDR (dB)	-13.0	1.8	1208
8	Relative Altitude Max LDR (m)	-109.8	70.7	1208
9	Minimum ρ_{HV}	0.85	0.05	403
10	Relative Altitude Min ρ_{HV}	-112.0	124.6	403

Table 3: Radar Estimated Melting Layer

Cooling Rates for Selected Cases

Day	Max Z_e (dBZ)	Rain Z_e (dBZ)	M (g/m^3)	$\Delta T/\Delta t$ (K/hr)
19	35	30	0.08	1.0
26	37	30	0.09	1.1
32	42	33	0.25	3.1
35	44	39	0.50	6.3
37	35	27	0.05	0.6
39	43	38	0.45	5.6
41	41	35	0.19	2.4
48	37	31	0.09	1.1
51	38	29	0.07	0.9
53	45	38	0.50	6.3

Figure Captions

Figure 1. Plots of nadir HH-VV measurements of bright band versus altitude for Day 39, 8 Feb, 1941 GMT.

Figure 2. Plots of nadir HH-HV measurements of bright band versus altitude for Day 53, Feb, 2043 GMT.

Figure 3. Scatter plot showing maximum bright band reflectivity versus altitude above mean sea level for Julian days 37 and 51.

Figure 4. ZHH measurements of bright band versus altitude for low reflectivity and high reflectivity cases on Day 51.

Figure 5. T-matrix calculations of prolate spheroid backscatter properties versus distance y as the particle melts. a) D_w equal 2 mm. b) D_w equal 5 mm.