P6R.5 SYSTEMATIC VARIATIONS OF RAINDROP SIZE SPECTRA WITH ALTITUDE DERIVED FROM WIND PROFILER: MEASUREMENTS FOR TRMM PR EVALUATION

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INTRODUCTION

Measurements of raindrop size distribution (DSD) are crucial for understanding rain formation processes in which the rainfall rate changes significantly, corresponding to the size spectrum (Atlas et al., 1999, 2000). Many processes affect DSD as drops fall, i.e., coalescence, nucleation, diffusion growth or evaporation, sedimentation, and breakup. Numerical modeling results show significant changes in the DSD during rainfall (e.g., List and McFarquhar 1990). However, few observations of the vertical profile of DSD are reported (Kobayashi and Adachi, 2001). More observations are needed.

This paper presents vertical variations of raindrop size distributions in precipitation occurred associated with typhoon 9707, on 20 June 1997 in Tukuba, Japan. A 404 MHz wind profiler (Fig.1) is used to derive DSD of precipitation at the Meteorological Research Institute, Tsukuba, Japan where is located about 100 km east of the typhoon center. The rain rate was ranging from 1 to 30 mm/h at the ground. It was continuously observed by the wind profiler. Table 1 summarizes the wind profiler system parameters

DSD RETRIEVAL FROM WIND PROFILER

UHF/VHF wind profilers detect signals scattered from turbulent fluctuations in the atmospheric refractive index, known as Bragg scattering (Williams et al. 1995). Received signals, which are Doppler shifted by air motion, are used to measure wind profiles. Under some conditions, the wind profiler can also detect signals from hydrometeors by Rayleigh scattering. A vertically pointing wind profiler can also measure the fall velocity of raindrops. Raindrop size can be derived from the fall velocity using a relationship between raindrop size and the terminal velocity. However, measured fall velocity combines the terminal velocity of raindrops and the ambient air motion. Small errors in the estimates of vertical air speed and turbulence can lead to significant errors in the derived DSD. Air motion should therefore be removed from the measured fall velocity.

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Doppler spectra measured with a wind profiler often have two distinct spectra, one from air motion and one from precipitation. Under such conditions, the wind profiler can measure both the fall velocity of raindrops and the ambient air motion, including the mean speed and the turbulence. The two distinct spectra facilitate the removal from the spectra of air motion effects.

In this paper, we have derived DSD by using a method developed by Kobayashi and Adachi (2005). The method is based on an iterative procedure and can derive arbitrarily shaped DSD, named ITRAN (ITerative Retrieval method for Arbitrarily shaped N (D): raindrop size distribution). This method assumes no prior shape of drop size distributions and fully derives automatically; additionally, it can be applied to large data volumes. Furthermore, it is insensitive to initial values. Here, we will briefly explain the retrieval method.

Fig.1 A photograph of the MRI wind profiler.

Table 1: MRI Wind profiler system parameters

<u>Frequency</u>: 404.0 MHz
<u>Power</u>: 35 kW (peak), 1.5 kW (average)
<u>Pulse width</u>: 1.67, 6.67 μseconds
<u>PRT:</u> 100 or 153.5 μseconds
<u>First range gate</u> 500 m (minimum)
<u>Antenna size</u>: 10.4X10.4 m (phased array)
<u>Antenna gain</u>: 33 dBi
<u>Beamwidth</u>: ≤ 4.1º
Beam steerability: 0 - 360º (azimuth), 0 - 15º (zenith)

The spectral density (S_{obs}) measured with a wind profiler is given by ,

$$S_{obs}(v) = [P_D S_D(v - v_0) * S_t(v) + P_t S_t(v - v_0)] * W, \quad (1)$$

where v is the Doppler velocity (upward +) and v_0 the mean vertical air velocity. P_t and P_D are the echo powers associated with refractive index irregularities and precipitation, respectively. S_t and S_D are the normalized spectral densities due to air turbulence and precipitation, respectively. W is the window function. The asterisk represents the convolution operation. The precipitation spectrum S_D is related to the raindrop size distribution $N(D)$ (D: raindrop diameter) as

$$S_D(v) = D^6 N(D) dD / dv / Z, \quad (2)$$

where Z is the reflectivity factor. Here we consider only spectral broadening due to turbulent motion. When $N(D)$ is derived, broadening effects in a measured Doppler spectrum should be removed by deconvolution.

The turbulent spectrum is assumed to have a Gaussian distribution, given by

$$S_t(v) = \frac{1}{(2\pi\sigma^2)^{1/2}} \exp(-v^2 / 2\sigma^2), \quad (4)$$

where σ is the spectral width. The Gaussian parameters are determined by non-linear least squares fitting using a Marquart method.

The turbulent peak velocity (v_{max}) is easily identified if the measured spectrum has two distinct peaks. The boundary of the two peaks (v_{min}) is the minimum position or is a few FFT points apart from the S_t peak. The spectrum ranging from v_{max}-v_{min} to v_{max}+v_{min} is fitted. If S_t is correctly positioned, the fitting works well.

Next, an iterative deconvolution of S_D*S_t is applied.

$$S_d^{i+1}(v) = < S_d^i(v) > \cdot \left\langle \frac{S(v)}{S^i(v)} \right\rangle_{i=0\ldots\infty}, \quad (5)$$

where X^i is the value of X (S_d or S) at the ^{i}th iteration. S_d^i is defined as P_D S_D in (1). $S^i(v)$ is the sum of $P_t S_t(v)$ and the convolution of $S_d^i * S_t$. The symbol < > denotes a running mean. The initial estimate S_d^0 is the observed spectrum smoothed by 1 ~ 5-points running mean, depending on the degree of statistical fluctuation. Multiplying S_d^i by the ratio of the observed spectrum S (v) to S^i(v) yields a new estimate of S_d^{i+1}. This process, by which new estimates of S_d^{i+1} are substitute in, is repeated. The precipitation spectrum is obtained point by point. Thus any shape of the DSD can be derived.

OBSERVATIONS

Corrected Doppler spectra measured with a 404 MHz profiler are available for 5 hours from 11:00 to 16: 00 JST at time intervals of 6 minutes and altitude intervals of 250 m. Doppler spectra of a low mode vertically pointing beam were averaged for one minute. Doppler spectra of a distinct peak were selected.

Here we will focus on a relationship between the rain water content (RWC) and the median volume diameter (D_0). Numerical models for the raindrop formation involving breakup and coalescence processes suggest that there should be correlations between the rain rate and the typical drop size. Many studies have reported the statistical relations between various integral parameters like the radar reflectivity factor, total number of drops (N_T), etc. derived from surface disdrometer measurements. However, no clear correlations have been reported between the rain rate and the median volume diameter.

Figure 2 shows wind profiler-derived D_0 versus the rain rate (R). The integral parameters were calculated from the derived DSD for diameters ranging from 0.5 to 5.5 mm to avoid from the effects of different drop size range because the minimum raindrop diameter of the derived DSD varied from 0.1 to 0.5, depending on atmospheric and rain conditions. Figure 2 shows no apparent correlation between R and D_0, as in studies by Bringi et al. (2002). Testud et al. (2001) described aircraft measurements in the Tropical Ocean and Global Atmosphere Coupled Ocean-Atmosphere Response Experiment and showed weak correlation for the entire dataset. He also showed an inverse relationship between N^* (intercept value of DSD) and D_0. During fall of raindrops, small and large drops, respectively, increase and decrease in number when collision-

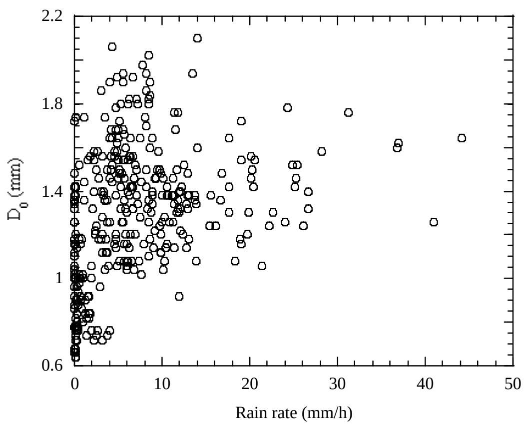

Fig. 2: Scatter plot of the wind profiler-derived median volume diameter versus the rain rate.

induced breakup occurs. These changes cause D_0 to decrease and N_T to increase. The opposite also occurs through the collision-induced coalescence. Therefore, an inverse relation between N_T and D_0 is expected. However, not even a weak relation between R to D_0 is present in Fig. 2 despite the apparent relation between R and N_T (not shown here). Median volume diameter characterizes the shape of DSD but is mathematically independent on N_T. Similar values of D_0 are observed during precipitation events with similar DSD and different R. This may be the reason of no clear correlation between R to D_0 shown in Fig.2.

Figure 3 shows a scatter plot of D_0 versus rain water content normalized by N_T. A clear relation is present, which is in agreement with results from Testud et al. (2001). The relationship between N_T and D_0 may be smeared when many events are considered. For a single precipitation event, N_T can be expected to relate to D_0.

Figures 4, 5, and 6 show vertical profiles of RWC, D_0 and N_T at 11:50, 11:57, and 13:09. The rain rates were 1, 6, and 10 mm/h, respectively at the ground. The values of RWC and N_T were normalized by the maximum values in the profile. The RWC profiles have peaks and dips at altitudes corresponding to those of the N_T. On the other hand, contrast to the uncorrelated results shown in Fig.2, clear inverse relations between D_0 and N_T appear. The N_T have peaks at 3.3 km at which the D_0 is minimum value in Fig.4. In Fig.5, a significant increase in the RWC at altitude of 2 km corresponds to a rapid decrease in D_0. Although most changes in RWC are in phase with change in N_T, there are some examples of inverse relationships between N_T and RWC. For example, at 4 km, RWC increases but N_T decreases in both figures 4 and 6. At this altitude, D_0 increases significantly, which leads to an increase in RWC. Fitting errors in

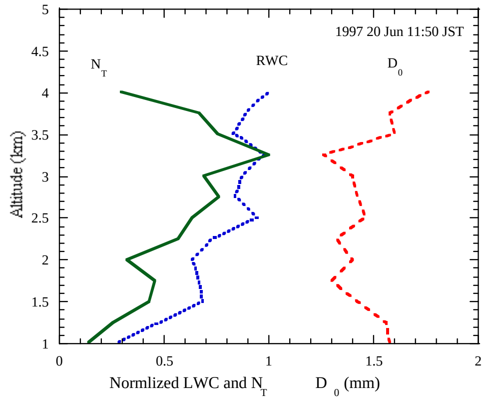

Fig.4 Vertical profile of RWC,D_0, and N_T at 11:50. The values of RWC and N_T are normalized by the maximum value in the profile. The inverse relationship between D_0, and N_T is clearly seen.

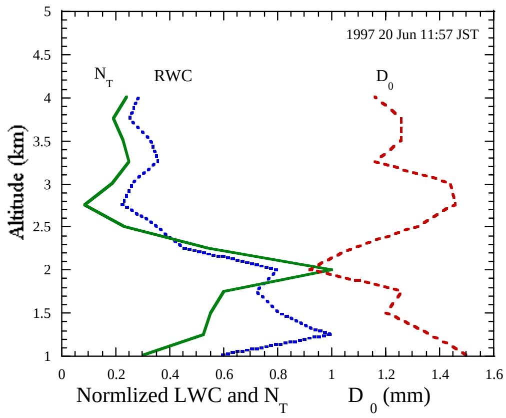

Fig.5 Same as Fig.4 except at 11:57.

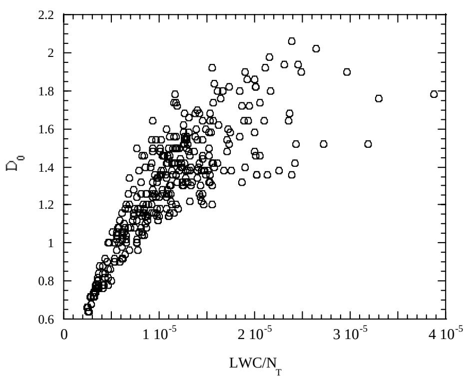

Fig.3 Scatter plot of D_0 and RWC/NT.

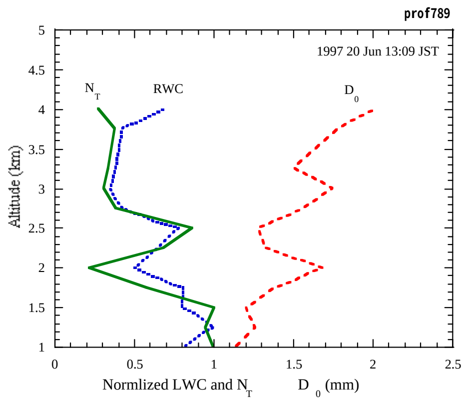

Fig.6 Same as Fig.4 except at 13:09.

the clear-air vertical motion overestimate D_0 and underestimate RWC and vice versa. However, D_0 is fairly insensitive to errors in v_0. The vertical air speed was almost zero. The vertical profile of Z is similar to that of RWC as estimated from the linear correlation. Errors in D_0 and RWC (N_T) are about 10 and 25percent even for an error of 0.5 m/s in v_0. The effects can be negligible associated with much larger variations of D_0 and N_T .

CONCLUTIONS

Precipitation associated with typhoon 9707 was observed by using a 404 MHz wind profiler on 20 June 1997 in Tsukuba, Japan. Vertical variations of the rain drop size distributions were derived by using an iterative retrieval method. This technique assumes no prior shape of DSD and therefore enables the vertical variations of D_0 as well as rain rate to be analyzed in detail. The median volume diameter is statistically independent of the RWC. The large variability of the total number of drops N_T between the rain events leads to the uncorrelated results. A vertical profile of each rain event shows a clear inverse relationship between N_T to D_0.

Whether these interesting results reported here are seen in other area or not is a problem to be solved. The precipitation radar (PR) onboard the Tropical Rainfall Measuring Mission (TRMM) has measured the vertical profile of rain rate since 1997. The rainfall rate is estimated from the radar reflectivity factor by using a power law: $R=aZ^b$ in which the parameter *a* and *b* are closely related to both N_T and D_0. Even for the current system, in which a single-frequency radar is operating, we can estimate raindrop size to some degree. For the Global Precipitation Mission (GPM), in which dual-frequency radar will be equipped, we can estimate the raindrop size more accurately, and therefore, we will examine the interesting results reported here for entire global scale.

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