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VERTICAL PROFILE OF RAINDROP SIZE DISTRIBUTION BY USING 400MHz WIND PROFILER IN STRATIFORM RAINFALL

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1. INTRODUCTION

The vertical profile of raindrop size distribution (DSD) is important for understanding rain formation processes. To improve the accuracy of rainfall intensity observed by precipitation radar and microwave radiometers carried on satellites, the information on the vertical profile of DSD is also important. Although there are some methods for measuring DSD on the ground, such as a disdorometer, a 2D-Video-Distrometer (2DVD), and so on, there is a few methods for measuring the vertical profile of DSD. A UHF/VHF wind profiler (WPR) is one of the most effective observation methods to provide the DSD profile.

In principle, if two peaks by air motion and by precipitation are appeared in the Doppler spectrum observed by WPR, a raindrop size can be derived from the fall velocity of precipitation using a relationship between raindrop size and terminal velocity. Several methods have been proposed to estimate DSD from WPR below the melting layer (e.g. Wakasugi et al., 1986; Sato et al., 1990; Kobayashi and Adachi, 2005).

The purpose of this study is to develop a database of DSD form rainfall (below the melting layer) to snowfall (above the melting layer) for the satellites precipitation retrieval by using a 400 MHz wind profiler (Adachi et al., 2001). The DSD is basically retrieved by the algorithm of Kobayashi and Adachi (2005) (hereafter Kobayashi algorithm). A new algorithm is proposed above the melting layer. To improve the accuracy of the DSD estimation, simultaneous observation data of the CRL Okinawa Bistatic Polarimetric Radar (COBRA) (Nakagawa et al., 2003) is used to calibrate the reflectivity of the 400 MHz WPR.

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2. OUTLINE OF OBSERVATIONS

Precipitation in Okinawa was observed in the 2004 rainy season during a field campaign called Okn-baiu04. This campaign was conducted as the part of the global satellite mapping of precipitation (GSMaP/CREST) project (Okamoto et al., 2004) and Lower Atomosphere and Precipitation Study (LAPS/CREST).

Figure 1 shows main observation sites of Okn-baiu04. The 400 MHz WPR is located at the National institute of Information and Communications Technology Ogimi (NiCT Ogimi) and COBRA at NiCT Nago

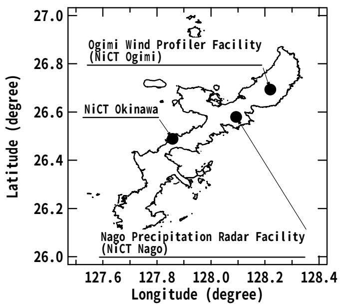

Fig.1 Location of the observation

2.1 Specifications of 400MHz WPR

The 400 MHz WPR provides vertical profiles of wind velocity vectors, turbulence and rainfall intensity by measuring Doppler frequency shift and echo intensity of radio signals scattered by turbulence and precipitation particles in the atmosphere. Table 1 shows the specifications of the 400 MHz WPR. This 400 MHz WPR uses active-phased array antenna with the aperture of 10.4m square. The WPR operated the pulse width of 1.33 and 2.0 μs for 5 directions toward zenith, north, south, east, and west (the zenith angle was 11 degrees for north, south, east and west beams) during the observation.

TABLE 1 Specifications of 400MHz WPR

400 MHz WPR	
Radar Type	Pulsed Doppler Rader
Frequency	443.0 MHz
Peak Power	20 kW
Average Power	2 kW
Pulse Length	1.33, 2.0, 4.0 μs
Pulse Repetition Frequency	6.25, 20 kHz
Antenna Type	24 × 24 element crossed array
Antenna Size	10.4 m × 10.4 m
Beam Width	3.3°
Beam Steerability	
Azimuth	0 - 360°
Zenith	0 - 15°
Observation Items	Verticul profiles of wind velocity, rainfall intensity, and virtual temperature

3. ESTIMATION OF DSD FROM DOPPLER SPECTRA

Figure 2 shows the outline of the estimation procedure of DSD from the Doppler spectrum. DSD below the melting layer is estimated by using the Kobayashi algorithm, while a new algorithm to estimate the vertical air motion is introduced above the melting layer in addition to the Kobayashi algorithm. The calibration factor for the 400 MHz WPR, which is obtained by a comparison with the radar reflectivity factor of COBRA, is used to finalize the DSD estimation.

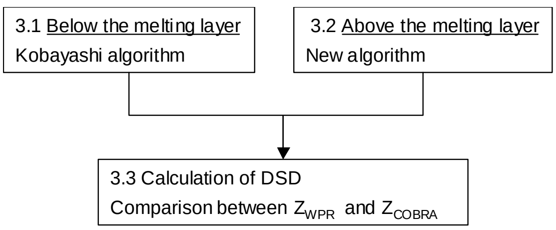

Fig.2 Procedure for estimating DSD from Doppler spectra

3.1 Algorithm below the melting layer

Two distinct peaks in the Doppler spectrum for air motion and precipitation are observed by the 400-MHz WPR (Fig. 4 (A)) in the most of stratiform rainfall. In the case of the Doppler spectra of air motion and precipitation can be divided from the observed Doppler spectrum, DSD is derived. The Kobayashi algorithm estimates DSD directly from the Doppler spectrum of precipitation based on an idea that observed Doppler spectrum can be expressed by the convolution of the Doppler spectrum of air motion and precipitation. In this algorithm the Doppler spectrum for air motion is assumed to have a Gaussian distribution. The spectral width and vertical air velocity of parameters of the Gaussian distribution are determined by the least square method. Figure 3 shows a flow chart of the Kobayashi algorithm.

The turbulence spectrum $S_t(v)$ is assumed to have a Gaussian distribution, which is given by

$$S_t(v) = \frac{1}{\sqrt{2\pi\sigma^2}} \exp\left[-\frac{(v - v0)^2}{2\sigma^2} \right] \qquad (1)$$

where σ is the spectral width. v is the Doppler velocity and $v0$ is the mean vertical air velocity.

The spectral density $S_{obs}(v)$ measured using a wind profiler is given by

$$S_{obs}(v) = P_t S_t(v) + S_D(v) * S_t(v) + P_n \qquad (2)$$

where $S_D(v)$ is the precipitation spectrum, P_t is the echo powers of air turbulence, P_n is the noise power, $*$ is the convolution.

The precipitation spectrum $S_D(v)$ is given by

$$S_D(v) = C \cdot N(D) \cdot D^6 \cdot \frac{dD}{dv} \qquad (3)$$

where C is *the* radar constant, and D is the raindrop diameter.

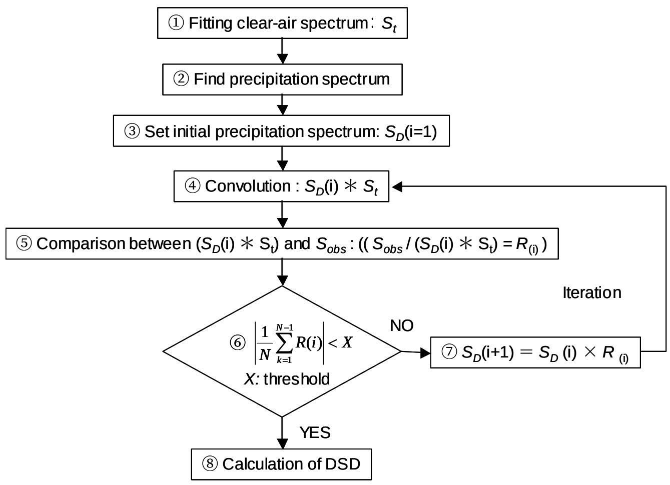

Fig.3 Flow chart for deriving DSD from Doppler spectra

3.2 Algorithm above the melting layer

When only small differences between air motion and precipitation fall velocity above the melting layer are observed in the Doppler spectrum in stratiform rainfall, the spectrum peak of air motion is difficult to find from the observed Doppler spectrum. However, when the Doppler spectrum for air motion can be found from the observed Doppler spectrum, DSD can be estimated using Kobayashi algorithm.

Figure 4(B) shows an example of the vertical profile of the observed Doppler spectrum above the melting layer. The Doppler spectrum with dotted line is the case in which both the air motion and precipitation are detected. The proposed algorithm in this study requires at least one reference data that is able to detect both the air motion and precipitation fall velocity in the vertical profile. We assume that the Doppler velocity at the peak of the precipitation spectra is uniform with heights above the melting layer, and the offset of the peak from the reference spectrum peak corresponds to additional air motion. Figure 5 schematically illustrates how to estimate the air motion above the melting layer. The Doppler spectrum with dashed line can detect the Doppler spectrum of the air motion $v0_b$ (□) and the Doppler velocity of the precipitation echo peak is expressed as Rv_b (●). At the top level in Fig. 5, the air motion is expressed as,

$$v0(h)=Rv(h)-Rv_b+v0_b \tag{4}$$

If the spectral width of vertical air motion in the Doppler spectrum (σ_b) is given by a fixed value of 0.1 m/s, DSD can be estimated by using the Kobayashi algorithm.

Figure 6 shows the flow chart of the method used to estimate the Doppler spectrum for air motion above the melting layer. The DSD is calculated assuming the relationship between the drop size D (mm) and the drop fall velocity v (m s^{-1}) given by

$$v = 0.837D^{0.142} \tag{5}$$

3.3 Calculation of DSD

In order to check the calibration of the 400 MHz WPR, the radar reflectivity factor of the 400 MHz WPR (referred to as Z_{WPR}) is compared the radar reflectivity factor of COBRA (referred to as Z_{COBRA}). Since the COBRA data was well calibrated by using active radar calibrator, the ratio between Z_{WPR} and Z_{COBRA} is used as the calibration factor for the DSD estimation. Figure 7 shows the flow chart of this procedure.

4. RESULTS

Stratiform rainfall from 15:00 to 24:00 on June 1, 2004 is analyzed. The rainfall type of this observation period is stratiform rainfall. We decide it from the COBRA data.

In this study, it is to estimate the vertical profile of DSD below and above the melting layer using the proposed algorithm. Figure 8 shows the vertical profile of vertical velocity and spectrum width at 12:36 UT. Figure 9 shows the vertical profile of raindrop size distribution by the

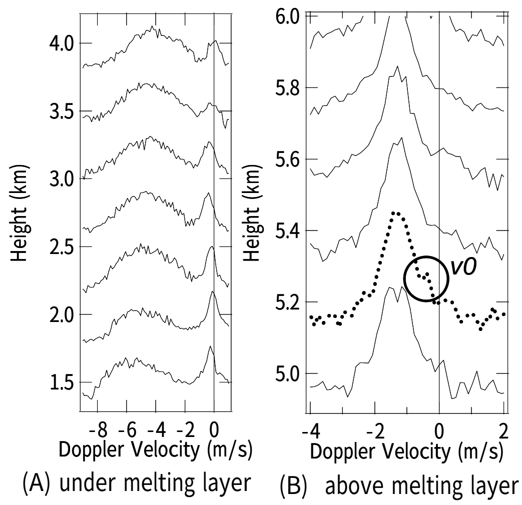

Fig.4 Vertical profiles of the Doppler spectrum observed by 400MHz WPR (2004/06/01 12:36 UTC)

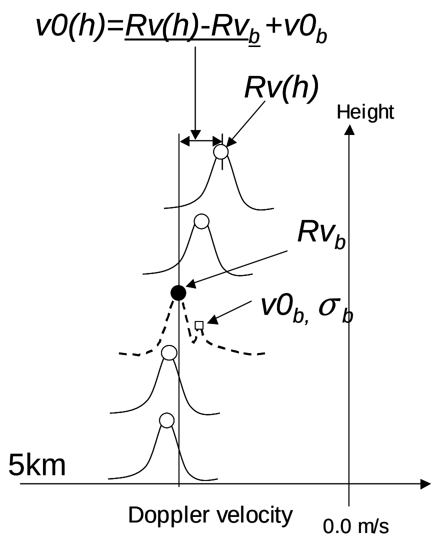

Fig.5 Method used to estimate air spectrum above the melting layer

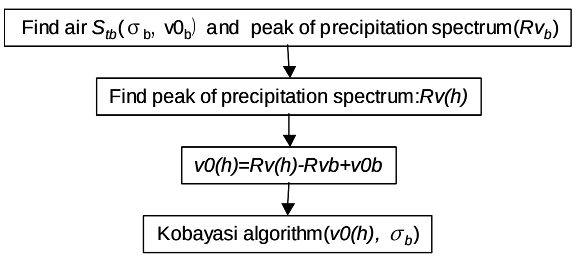

Fig.6 Flow chart for estimating air spectrum above the melting layer

proposed algorithm. The shape of DSD above the melting layer almost does not vary with heights (Fig.9 (a)). The trend vertical variation of DSD does small change below the melting layer. There is a small number of raindrops and the maximum diameter of raindrops ranged from 3 to 5 mm below the melting layer (Fig.9 (b)). The number of raindrops and maximum diameter of raindrops are large near the ground. This analytic result is the first step.

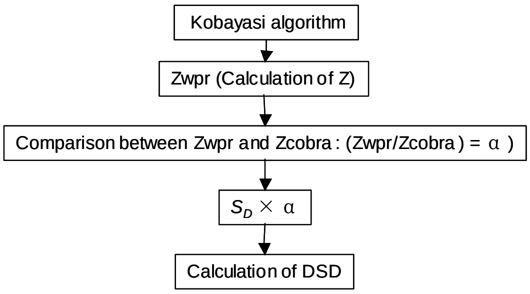

Fig. 7 Flow chart for calculating DSD

In the future, we will analyze other cases in stratiform rainfall by the proposed algorithm, and compare the result with the numerical model above the melting layer, and compare the result with DSD on the ground using 2DVD and disdrometer. We will also analyze convective rainfall.

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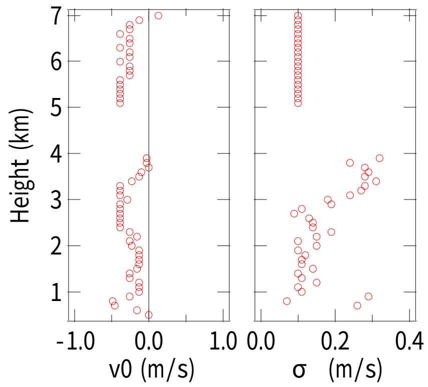
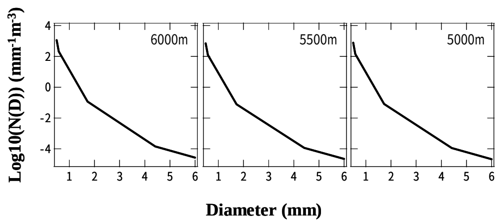
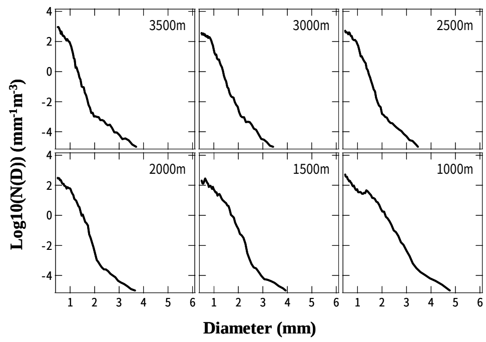


Fig.8 Vertical profile of vertical velocity and spectrum width (2004/06/01 12:36 UTC)



(a) Above melting layer



(b) Below melting layer

Fig.9 Vertical profile of raindrop size distribution (2004/06/01 12:36 UTC)

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