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A METHOD TO ANALYZE THE METEOROLOGICAL RADAR ECHO OBSERVED AT SYOWA STATION, ANTARCTICA

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Abstract: In this paper, we propose an approach to evaluate precipitation rate from meteorological radar echo data and snow particle size distribution data. We applied our approach to analyze two sets of radar echo data which were observed at Syowa Station, Antarctica on April 6 and October 1, 1988. Our evaluated precipitation rates agree with the results evaluated from the Z factor—precipitation rate relations, which were measured at Syowa Station, Antarctica. On the bases of our results, we have confidence that our approach can be applied to analyze the meteorological radar echo in Antarctica.

1. Introduction

In February 1988, a vertically pointing meteorological radar (9.41 GHz) was installed at Syowa Station, Antarctica. Observations of precipitation using this radar system were carried out in 1988 and 1989, and results of these observations have been reported by WADA (1990) and WADA and KONISHI (1992). In these reports, observed radar echo data were analyzed by using Z factor—precipitation rate relations, which had been observed at Syowa Station in 1989 intermittently (KONISHI *et al.*, 1992).

In this paper, we determined an approach to analyze meteorological radar echo data which is based on the calculation of back scattering cross section using a snow particle model and observed snow particle size distribution function. We evaluate precipitation rates at each altitude using this method.

2. Meteorological Radar Equation

The meteorological radar equation, which ignored two-way attenuation due to precipitation, is expressed by the radar reflectivity factor (Z factor) as follows (SKOLNIK, 1980):

$$P_{\mathrm{r}} = \frac{C}{R^2}\sum_i \sigma_i = \frac{C}{R^2}\,\frac{\pi^5}{\lambda^4}\left|\frac{\varepsilon-1}{\varepsilon+2}\right|^2 Z, \tag{1}$$

where P_{r} is the received power, C is a constant depending on radar system parameters such as transmitted power, R is distance from radar to target, $\sum_i \sigma_i$ is the average total back scattering cross section per unit of volume, λ is wavelength and ε is dielectric constant of precipitation particles. If we assume that precipitation particles are spherical dielectric bodies, $\sum_i \sigma_i$ is expressed by Mie's back scattering cross section of a particle $Q_\pi(r)$ and particle size distribution function $N(r)$.

$$\sum_i \sigma_i = \beta = \int N(r)\, \mathrm{Q}_\pi(r)\, \mathrm{d}r, \tag{2}$$

where r is the particle radius and β is total back scattering cross section per unit of volume. This $\mathrm{Q}_\pi(r)$ is calculated from the following formula (KERR, 1964).

$$\mathrm{Q}_\pi = \frac{\lambda^2}{4\pi} \left|\sum_{n=1}^{\infty} (-1)^n (2n+1)(a_n+b_n)\right|^2,$$

$$a_n = -\frac{j_n(\rho)[m\rho j_n(m\rho)]' - j_n(m\rho)[\rho j_n(\rho)]'}{h_n^{(2)}(\rho)[m\rho j_n(m\rho)]' - j_n(m\rho)[\rho h_n^{(2)}(\rho)]'},$$

$$b_n = -\frac{j_n(\rho)[m\rho j_n(m\rho)]' - m^2 j_n(m\rho)[\rho j_n(\rho)]}{h_n^{(2)}(\rho)[m\rho j_n(m\rho)]' - m^2 j_n(m\rho)[\rho h_n^{(2)}(\rho)]'}, \tag{3}$$

where $\rho = 2\pi r/\lambda$, $m(=\sqrt{\varepsilon})$ is the complex refractive index, j_n are spherical Bessel functions of the first kind, $h_n^{(2)}$ are spherical Hankel functions of the second kind, and the primes indicate differentiation with respect to the argument.

To evaluate precipitation rate P [mm/h], we introduce reflection coefficient per unit precipitation rate β_0.

$$\beta = P \cdot \beta_0 \ [\mathrm{cm^2/m^3}],$$

$$\beta_0 = \frac{\int N(r)\, \mathrm{Q}_\pi(r)\mathrm{d}r}{15.08 \int r^3\, N(r)\, v_s \rho_s \mathrm{d}r} \ [\mathrm{cm^2/m^3/mm/h}], \tag{4}$$

where v_s [m/s] and ρ_s[g/cm^3] are falling speed and density of the precipitation particles. Substituting eqs. (2) and (4) into eq. (1), the precipitation rate P is expressed as follows:

$$P = \frac{\pi^5}{\lambda^4} \left|\frac{\varepsilon-1}{\varepsilon+2}\right|^2 \frac{Z \times 10^{-6}}{\beta_0} \ [\mathrm{mm/h}]. \tag{5}$$

3. Characteristics of Snow Particles

In eq. (5), the wavelength λ is known and the Z factor is obtained from radar echo. To calculate β_0, we are interested to know the snow particle size distribution function $N(r)$, complex dielectric constant ε_s, falling speed v_s and density ρ_s of the snow particles.

HATANAKA et al. (1993) obtained the size distribution functions $N(r)$ on the ground from snow particle VTR images which had been recorded at Syowa Station, Antarctica. The size distribution function at each altitude is probably

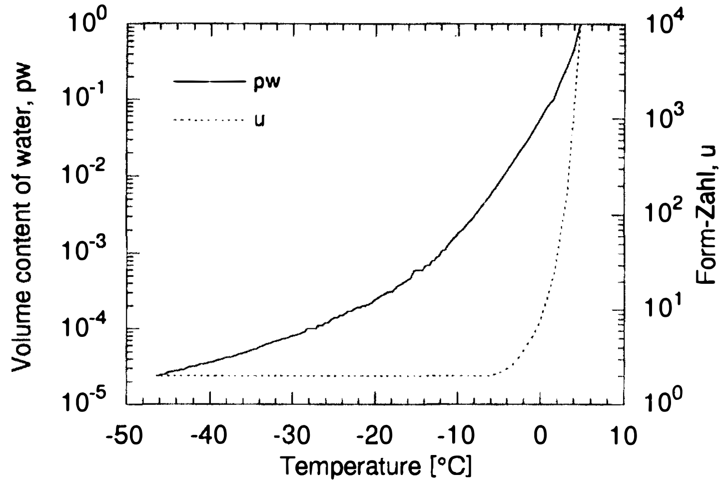

Fig. 1. The empirical relation between the volume content of water p_w, "Form-zahl" u of a snow particle and the temperature of the snow particle (NISHITSUJI et al., 1983).

different from that on the ground, but it is very difficult to estimate the size of distribution at each altitude. In this analysis, we assume that the snow particle size distribution function at each altitude is the same as the distribution observed on the ground.

To estimate the dielectric constant of the snow particle ε_s, we used Nishitsuji's snow particle model (NISHITSUJI, 1971; AWAKA *et al.*, 1985). In this model, a snow particle is considered to be a mixture of air, ice and water, and its dielectric constant is given by the following empirical formula:

$$\frac{\varepsilon_s-1}{\varepsilon_s+u} \fallingdotseq p_w \frac{\varepsilon_w-1}{\varepsilon_w+u} + \frac{\sqrt{p_w}-p_w}{0.92} \frac{\varepsilon_i-1}{\varepsilon_i+u}, \tag{6}$$

where ε_w and ε_i are the complex dielectric constants of water and ice, u is the "Form-zahl", and p_w is the volume content of water. When the temperature of the particle is given, the values of ε_w and ε_i are uniquely determined by RAY (1972). The values of p_w and u are changed depending on the state of snow-falling, *e.g.* dry-, moist-, wet- and watery-snow. In this model, it is assumed that the state of snow is mainly determined by the temperature as shown in Fig. 1 (NISHITSUJI *et al.*, 1983). The density of the particle ρ_s [g/cm³] is estimated from p_w using the following empirical relation (NISHITSUJI, 1971).

$$p_w \fallingdotseq \rho_s^2 . \tag{7}$$

The falling speed of the particle v_s [m/s] is computed from the following formula (MAGONO and NAKAMURA, 1965).

$$\begin{aligned} v_s &= 8.8\,[2r_s(\rho_s-\rho_a)]^{1/2} \quad \text{[m/s]} \quad (0.05 \le \rho_s \le 0.3 \text{ [g/cm}^3\text{]}) \\ v_s &= 3.3\,(\rho_s-\rho_a)^{1/2} \quad \text{[m/s]} \quad (\rho_s < 0.05 \text{ [g/cm}^3\text{]}), \end{aligned} \tag{8}$$

where ρ_a[g/cm^3] is density of air.

In this model, every characteristic of a snow particle except for the size distribution function is determined by the temperature of the particle. We further assume that the snow particles are in thermal equilibrium with the surrounding atmosphere, and we use the atmospheric temperature data at each altitude measured by radiosonde.

4. Various Data Used in This Analysis

In this paper, we analyze two sets of radar echo data observed at Syowa

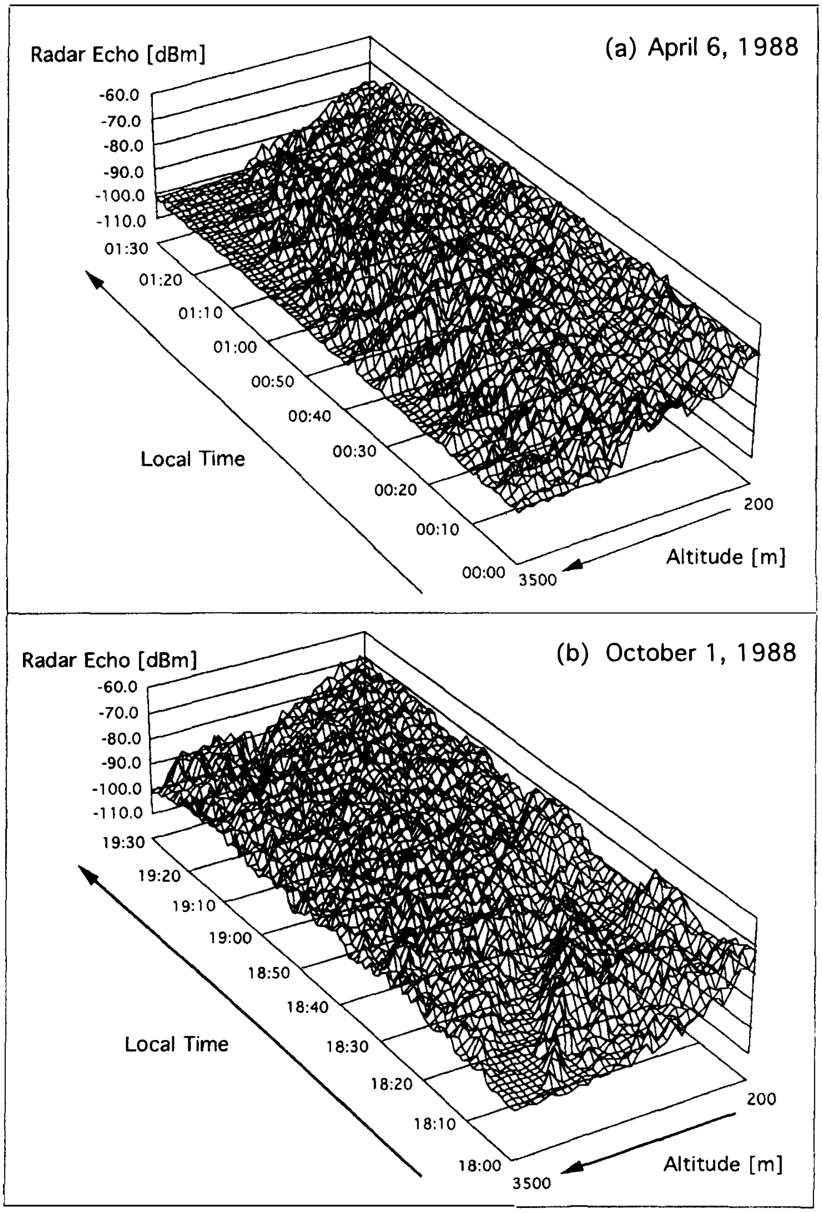

Fig. 2. Radar echo data observed at Syowa Station, Antarctica. (a) From 0000 LT to 0130 LT on April 6, 1988. (b) From 1800 LT to 1930 LT on October 1, 1988.

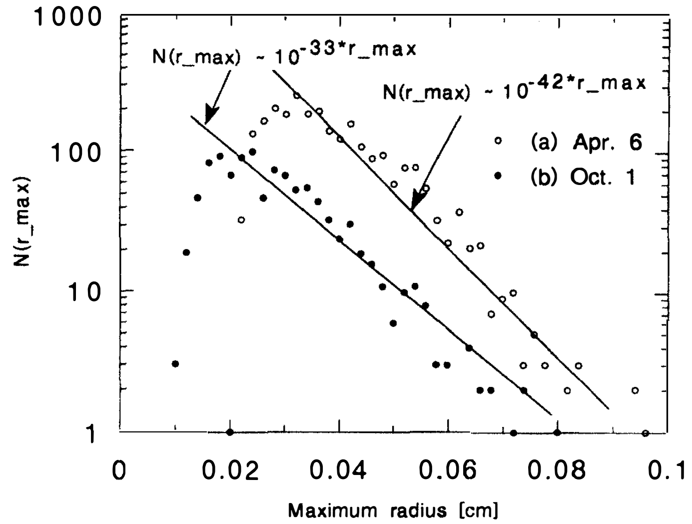

Fig. 3. Averaged snow particle size distribution function N(r_max) based on the maximum radius of particles r_max at Syowa Station obtained from two sets of VTR images, which had been recorded (a) from 2113 LT on April 5 to 0133 LT on April 6 and (b) from 1800 LT to 1944 LT on October 1, 1988 (HATANAKA et al., 1993).

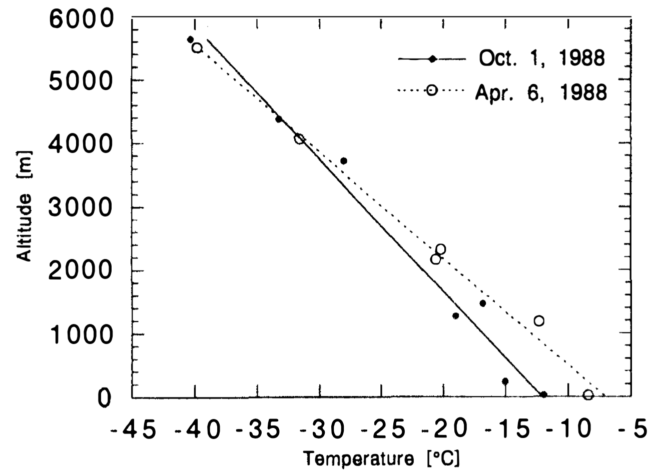

Fig. 4. Atmosphere temperature data above Syowa Station, measured by radiosonde at 0200 LT on April 6 and 1400 LT on October 1, 1988.

Station, Antarctica from 0000 LT to 0130 LT on April 6 (case a) and from 1830 LT to 1930 LT on October 1 (case b), 1988 as shown in Figs. 2a and 2b. The meteorological conditions of cases a and b are summarized in Table 1. Corresponding snow particle size distribution functions based on maximum radius of the particles (HATANAKA *et al.*, 1993) and the temperature data measured by radio-sonde are shown in Figs. 3 and 4, respectively. Figure 5 shows examples of complex dielectric constant of snow particles, which are calculated from eq. (6)

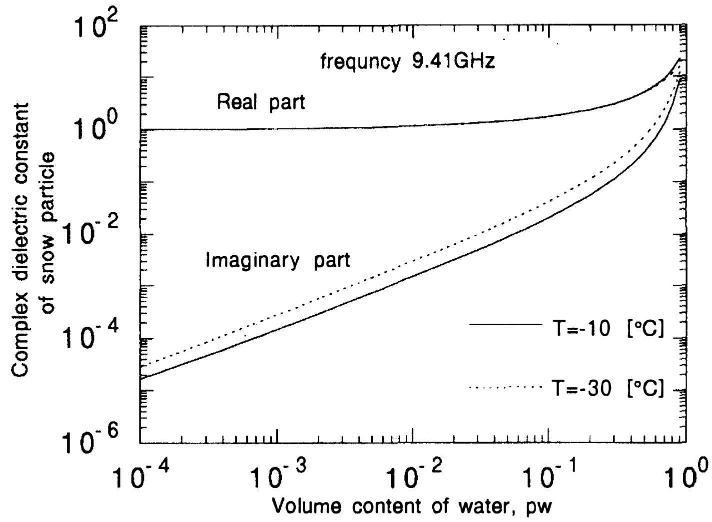

Fig. 5. Examples of calculated complex dielectric constant of snow particles.

using the values of u, volume content of water p_w in Fig. 1 and the values of ε_w and ε_i of RAY (1972).

5. Results and Discussion

Examples of obtained precipitation rate at each altitude, which are evaluated from the echo data in Figs. 2a and 2b using eq. (5), are shown in Figs. 6a and 6b. In these figures, we plot the results, which are evaluated from the same echo data in Figs. 2a and 2b using the following measured Z factor—precipitation rate relations at Syowa Station in 1989 (KONISHI *et al.*, 1992).

$$Z=a^*P^b. \tag{9}$$

Case A: $a=74$, $b=1.4$,
Case B: $a=104$, $b=1.3$,
Case C: $a=10$, $b=1.2$.

The meteorological conditions, when KONISHI *et al.* measured these relations, are summarized in Table 1.

In Figs. 6a and 6b, our results using eq. (5) agree with those of eq. (9) in cases A and B, but the result in case C is much higher than the others. We think that this difference is mainly caused by the difference of the particle size distribution, because the particles in case C were concentrated in a narrow size range, up to 3 mm in diameter (KONISHI *et al.*, 1992). We compare our result and those using eq. (9) in cases A and B more precisely. At lower altitude, our result fits case B, but it fits case A at higher altitude. This means that in our analysis, at higher altitude, the temperature of particles is lower, as shown in Fig. 4; the imaginary component of dielectric constant of the snow particles is

Table 1. Meteorological conditions for five snowfalls.

	Date	t (°C)	w_s (m/s)	Max-r (cm)	References
Case A	October 24–25, 1989	−3	2	0.3	KONISHI *et al.*, 1992
Case B	July 1, 1989	−9	2	0.5	KONISHI *et al.*, 1992
Case C	May 21–22, 1989	−14	3	0.15	KONISHI *et al.*, 1992
Case a	April 6, 1988	−8.1	2.8	0.096	HATANAKA *et al.*, 1993
Case b	October 1, 1988	−13.5	0.7	0.081	HATANAKA *et al.*, 1993

t: Mean surface air temperature. w_s: Mean wind speed. Max-r: Maximum radius of snow particles.

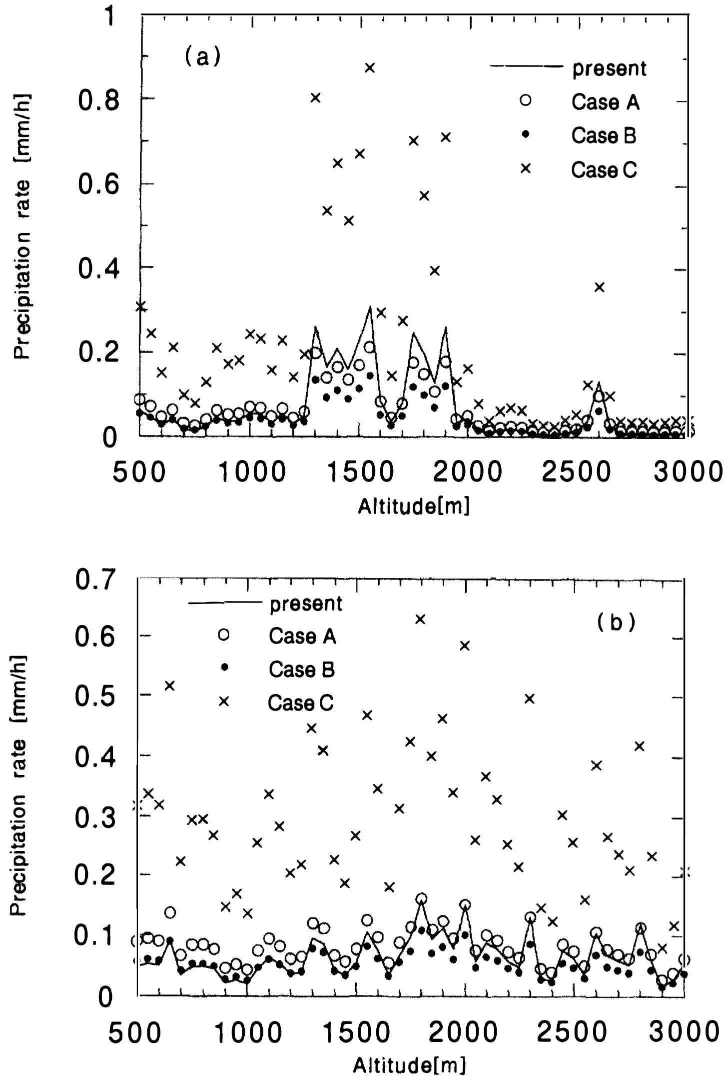

Fig. 6. Evaluated precipitation rates at altitudes from 500 m to 3000 m. The line is the result obtained using eq. (5); cases A, B and C are based on the Z factor—precipitation rate relations (KONISHI et al., 1992). (a) At 0000 LT on April 6, and (b) at 1900 LT on October 1, 1988.

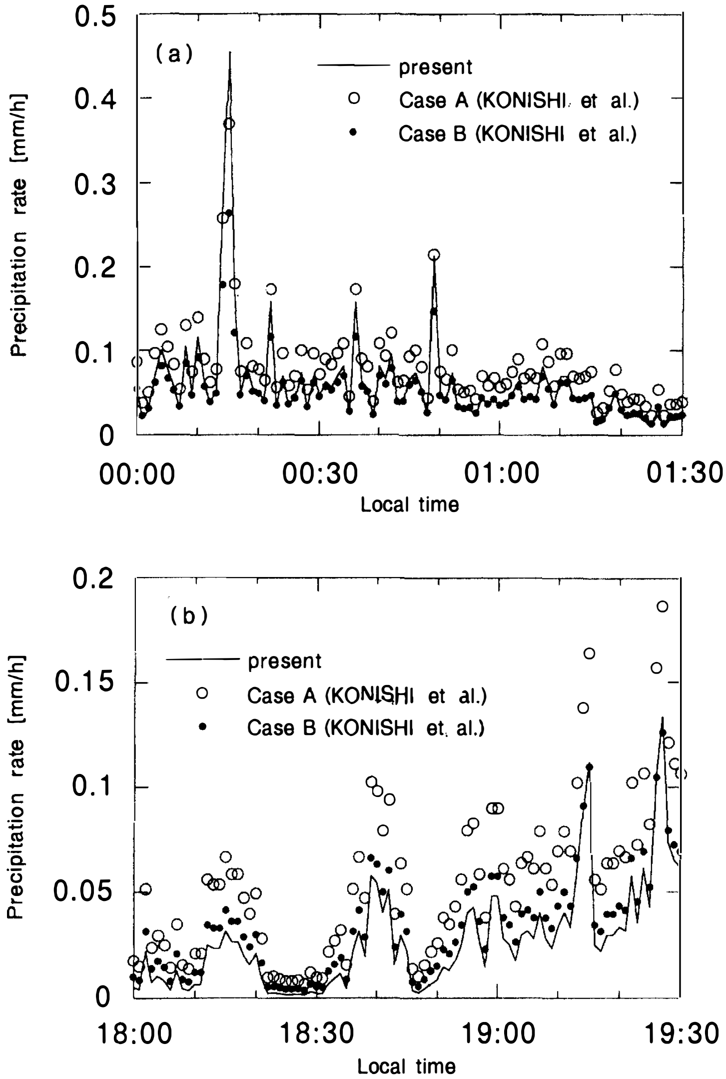

Fig. 7. Evaluated time variation of precipitation rate at altitude 500 m. The line is obtained using eq. (5), and cases A, B and C are based on the Z factor—precipitation rate relations (KONISHI et al., 1992). (a) From 0000 LT to 0100 LT on April 6, and (b) from 1830 LT to 1930 LT on October 1, 1988.

decreased, as shown in Fig. 5; and the reflection coefficient per unit precipitation rate β_0 is decreased. This effect corresponds to decreasing coefficient a in eq. (9).

Figures 7a and 7b show the time variation of precipitation rate at the altitude of 500 m on April 6 and October 1, 1988. In these figures, our results fit the results using eq. (9) in case B.

6. Concluding Remarks

We proposed an approach to evaluate precipitation rate from the meteorological radar echo data and the snow particle size distribution functions, and we analyzed two cases of radar echo data which had been observed at Syowa

Station, Antarctica on April 6 and October 1, 1988. Our obtained precipitation rates agreed with the results evaluated by the measured Z factor—precipitation rate relations (KONISHI *et al.*, 1992). According to this study, Nishitsuji's snow particle model, which is one of the basic assumptions in this approach, can be applied to analyze radar echo data in Antarctica.

Since MURAMOTO *et al.* (1992) and HATANAKA *et al.* (1993) showed that snow particle size distribution function can be obtained using a CCD camera system and a personal computer, we think that this method is one of practical approaches for evaluating annual variation of snowfall in Antarctica automatically, using a meteorological radar system.

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