Retrieval of Raindrop Size Distribution, Vertical Air Velocity and Water Vapor Attenuation Using Dual-Wavelength Doppler Radar Observations

Lin Tian¹, Gerald M. Heymsfield², Lihua Li¹, and Ramesh C. Srivastava³

¹Goddard Earth Science and Technology Center, University of Maryland, Baltimore, Maryland
²NASA Goddard Space Flight Center, Greenbelt, Maryland
³Department of the Geophysical Sciences, The University of Chicago, Chicago, Illinois

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Corresponding author’s address: Lin Tian, NASA Goddard Space Flight Center, Code 613.1, Building 33, Rm C416, Greenbelt, MD 20771.
Email: tian@agnes.gsfc.nasa.gov
Abstract

Two techniques for retrieving the slope and intercept parameters of an assumed exponential raindrop size distribution (RSD), vertical air velocity, and attenuation by precipitation and water vapor in light stratiform rain using observations by airborne, nadir looking dual-wavelength (X-band, 3.2 cm and W-band, 3.2 mm) radars are presented. In both techniques, the slope parameter of the RSD and the vertical air velocity are retrieved using only the mean Doppler velocities at the two wavelengths. In the first method, the intercept of the RSD is estimated from the observed reflectivity at the longer wavelength assuming no attenuation at that wavelength. The attenuation of the shorter wavelength radiation by precipitation and water vapor are retrieved using the observed reflectivity at the shorter wavelength. In the second technique, it is assumed that the longer wavelength suffers attenuation only in the melting band. Then, assuming a distribution of water vapor, the melting band attenuation at both wavelengths and the rain attenuation at the shorter wavelength are retrieved. Results of the retrievals are discussed and several physically meaningful results are presented.
Measuring global rainfall property by satellite such as Tropical Rain Measuring Mission (TRMM) is important for understanding weather and how to improve the weather prediction. Rain has been very difficult to measure accurately over a wide range of intensities because there is generally insufficient information about the rain properties. A dual-wavelength radar, such as the one proposed for the Global Precipitation Measurement (GPM) can improve the accuracy of rainfall estimation. However, the 14 and 35 GHZ frequencies proposed for the GPM radar may not resolve the raindrop size distribution in light rain due to the small difference of the reflectivity at the two frequencies. By using combination of 94 GHz and 10 GHz reflectivity, and Doppler information that is not available on GPM, it has been found in this paper that light rain can be estimated more accurately.

A method is developed for estimating rainfall rate and other parameters in light stratiform rain using airborne Doppler radar observations at 94 and 10 GHz. The method uses the difference of the return power as well as Doppler velocity measured at the two frequencies to estimate rainfall rate, vertical air velocity, and water vapor attenuation that is nearly proportional to the water vapor content. The magnitudes and vertical variation of the averaged velocity are remarkably similar to what has been reported in the literature for area-averaged vertical velocity in extensive stratiform rain. The averaged profile of rainfall rate, and raindrop size distribution has shown evidence of drop breakup below the melting band and evaporation in the lower levels in the region of downdrafts. The potential ability to retrieve raindrop size distribution, vertical air velocity, and water vapor distribution as functions of the height is exciting because it offers the possibility of direct calculation of profiles of evaporation and, therefore, latent heating and cooling in the atmosphere using microphysical principles.
1. Introduction

Over the past several years, the Tropical Rainfall Measuring Mission (TRMM) has provided data for understanding storm formation and structure and improving weather prediction. Measurements by the single wavelength TRMM precipitation radar (PR) have been used to estimate profiles of attenuation-corrected reflectivity and rainfall rate (RR) (Iguchi et al. 2000). The data have also been used in conjunction with numerical cloud models to estimate latent heating/cooling.

The retrieval of RR by the TRMM PR and other single wavelength radars, is crucially dependent on the raindrop size distribution (RSD). Uncertainties in RSD are an inherent source of error in estimated RR. If the RSD depended on a single parameter, then the reflectivity (Z) at a single non-attenuating wavelength would suffice for retrieving RR. A commonly used one-parameter RSD is the Marshall-Palmer (MP) distribution (1948):

\[ N(D) = N_0 \exp(-\Lambda D) \]  

(1.1).

Here \( N(D) \) is the concentration of drops of diameter \( D \) per unit diameter interval, \( N_0 \) is a fixed intercept (\( = 8 \times 10^{-2} \text{cm}^{-4} \)) and \( \Lambda \), the slope of the distribution, is the sole variable parameter. Marshall and Palmer proposed the exponential distribution based on ground measurements of RSD in temperate latitudes. It has been widely used for the estimation of RR by radar. Subsequent measurements have confirmed the exponential form of the RSD, especially for stratiform rain, but have found the \( N_0 \) to be highly variable. For example, Waldvogel (1974) found that while the RSD is well represented by (1.1), \( N_0 \) ranges 2-3 orders of magnitude. This variability is a major source of error in estimation of RR by single-wavelength radar. Other measurements of RSD, especially in convective rain, have shown that it is more accurately described by the gamma distribution (Ulbrich 1983) which has three variable parameters. In this
paper, we shall confine ourselves to the two-parameter exponential RSD (1.1) with variable $N_0$ and $\Lambda$ since we shall be concerned only with stratiform rain.

A dual-wavelength radar, such as the one proposed for the Global Precipitation Measurement (GPM) can provide RSD information and hence improve the accuracy of rainfall estimation. However, the 14 and 35 GHz frequencies proposed for the GPM radar may not resolve the RSD in light rain due to the small difference of the reflectivity at the two frequencies. A combination of mm and cm wavelengths may provide more accurate information for light rain. Moreover, Doppler velocity, which improves the RSD estimation, as shown in this paper (see also, Meneghini et al. 2003), will not be available for the proposed GPM radar.

This paper presents techniques for estimating RSD and other parameters in light stratiform rain using Doppler radar observations at 3.2 cm and 3.2 mm wavelengths. In section 2, we present the basic radar equations. In section 3 we give a brief description of the radar observations used in this study. In sections 4 and 5, we present two techniques for retrieving the parameters of the RSD (1.1), the melting band attenuation, and water vapor attenuation and present the results obtained by application of the methods to the radar observations. Section 6 presents a summary and discussion of the work.

2. Basic Equations

This section presents the basic equations and introduces our notation and terminology. We neglect signal fluctuations, assume identical radar beams and simultaneous observations at the two wavelengths.

The received power, at range $r$, is expressed by the ‘measured’ equivalent radar reflectivity factor, $Z^{(m)}(r)$, or simply measured reflectivity; it is given by:

$$Z^{(m)}_{cm}(r) = Z_{cm} (r) \times A_{cm} (0, r)$$  \hspace{1cm} (2.1)
where $Z_{cm}(r)$ is the 'true' reflectivity and $A_{cm}(0, r) \leq 1 \) is the attenuation factor, which accounts for losses due to the medium between the radar at $r = 0$ and the range $r$. (A list of symbols and units is given in Appendix A.) In the above and subsequent equations, the subscripts $mm$ and $cm$ refer to the two wavelengths, 3.2 $mm$ and 3.2 $cm$ respectively, and the superscript $(m)$ refers to measured quantities. The reflectivity that would be measured in the absence of attenuation, so-called 'true' reflectivity is given by:

$$Z_{cm} = c_2 \int_{D_{min}}^{D_{max}} N(D) \sigma_{BACK, cm}(D) dD$$

where $c_2$ is a numerical unit conversion factor and $\sigma_{BACK}(D)$ is the back-scattering cross-section of drop of diameter $D$. The integral is taken from a minimum diameter $D_{min}$ to a maximum diameter $D_{max}$. Similar equations hold for the $mm$ wavelength. The reflectivity is often expressed in $dBZ$ units. We shall use $\xi$ to represent $Z$ in $dBZ$ unit (we use corresponding Greek letters to express quantities in $dB$ units). Equation (2.1) may be written as:

$$\xi^{(m)}(r) = \xi_{cm}(r) - \alpha_{cm}(0, r)$$

where

$$\xi_{cm} = 10 \log_{10} (Z_{cm}), \quad \xi^{(m)} = 10 \log_{10} (Z^{(m)}_{cm})$$

and

$$\alpha_{cm}(0, r) = -10 \log_{10} [A_{cm}(0, r)]$$
Here $\alpha_{cm}(0,r)$ is the two-way attenuation in $dB$ between the radar and the range $r$. We decompose the attenuation factor into losses due to gases and precipitation and express it in terms of specific attenuation:

$$A(0,r) = A^{(gas)}(0,r) \times A^{(ppt)}(0,r) = \exp\left\{-\int_0^r [k^{(gas)}(r') + k^{(ppt)}(r')] dr'\right\}$$  \hspace{1cm} (2.6).

In the above, wavelength subscripts have been omitted and $k^{(gas)}$ and $k^{(ppt)}$ are the two-way specific attenuation coefficients due to gases and precipitation, respectively. We have not included attenuation by cloud water because we shall apply our method to stratiform rain with melting band and cloud water is not expected in the rain region. The two-way specific attenuation coefficient by rain is given by:

$$\kappa = c_k \int_{d_{min}}^{d_{max}} N(D) \sigma_{EXT}(D) dD$$ \hspace{1cm} (2.7),

where $\kappa$ is $k$ expressed in $dB/length$ units, $\sigma_{EXT}(D)$ is the extinction cross-section of drop of diameter $D$ and $c_k$ is a numerical unit conversion factor.

The measured mean Doppler velocity at the cm wavelength, $\overline{v}_{cm}^{(m)}$, is the reflectivity-weighted mean of the vertical velocity of the drops. It is given by:

$$\overline{v}_{cm}^{(m)} = \overline{v}_{cm} - w$$ \hspace{1cm} (2.8),

and
\[
\bar{V}_c = \frac{\int_{D_{\text{min}}}^{D_{\text{max}}} N(D) \sigma_{\text{BACK, cm}}(D) V_i(D) dD}{\int_{D_{\text{min}}}^{D_{\text{max}}} N(D) \sigma_{\text{BACK, cm}}(D) dD}
\]

where \( \bar{V} \) is the reflectivity-weighted mean terminal fall speed, \( w \) is the vertical air velocity, and \( V_i(D) \) is the terminal fall speed of drop of diameter \( D \). The vertical air velocity is taken positive upwards, while the mean Doppler velocity and the terminal fall speed are taken positive downwards. Similar equations hold for the \( mm \) wavelength.

3. Observations

The observations used in this study were made in stratiform rain in the Florida area using nadir-looking airborne X-band (3.2 cm) and W-band (3.2 mm) Doppler radars. The aircraft flies at an altitude of 20 km. The range resolution for both wavelengths is 37.5 m and the beamwidths are 2.9°X2.9° and 0.6°X0.8° giving ground-footprints of approximately 1.0X1.0 and 0.21X0.28 km for the cm and mm wavelengths, respectively. The characteristics of the radars are summarized in Table 1. To minimize errors due to unmatched beams and signal fluctuations, the observations are averaged along the flight line. The Doppler velocity is corrected for aircraft motion and folding. Heymsfield et al. (1996) and Li et al. (2004) give detailed descriptions of the radars and data reduction procedures.

Figures 1 and 2 show the flight path and a nearby sounding. Figure 3 shows reflectivities and mean Doppler velocities for an extensive stratiform rain system with the top of the melting band at a height of about 4.3 km. The large decrease in reflectivity at the shorter wavelength with distance downwards below the melting band indicates severe attenuation by rain. Indeed, at distances less than about 135 km and greater than about 175 km, the \( \xi_{nm}^{(m)} \) is below the radar.
noise level near the surface. At the longer wavelength, we see a number of precipitation trails that can be followed from the ice into the rain region, and increases in the melting band reflectivity where the trails intersect the melting band. About eight trails can be identified at distances of about 121, 125, 132, 138-148, 150-156, 158-162, 163-165, 171 and 175 km; these distances refer to the points where the trails intersect the melting band. The trails can also be seen in the field of $\tilde{V}_\text{cm}^{(m)}$. From the shapes of the trails, it is clear that the winds must be varying with height in both speed and direction. Unfortunately, sufficient nearby wind data were not available to perform calculations of trail shape. But it is clear that caution must be exercised in interpreting the vertical distributions of retrieved parameters in terms of column models of RSD evolution.

4. Retrieval I: Method and Results

Retrieval method I involves three steps. In the first step we use only the measured mean Doppler velocities to retrieve $\Lambda$, $D_0$ and $w$. In the second step we use an additional parameter, namely, $Z_{\text{cm}}^{(m)}$, to retrieve $N_0$. In the third and final step, we use $Z_{\text{mm}}^{(m)}$ to retrieve the distribution of water vapor attenuation. In the second and third steps, it is necessary to assume that there is no attenuation at the cm wavelength; however, the retrieval of $\Lambda$, $D_0$ and $w$ in the first step is independent of attenuation at both wavelengths.

a. Step 1 – Retrieval of $\Lambda$, $D_0$ and $w$.

(i) Method

The mean Doppler velocities are independent of attenuation and $N_0$ and, moreover, their difference is independent of $w$:
\[ \partial \bar{V}^{(m)} = \bar{V}^{(m)}_{cm} - \bar{V}^{(m)}_{mm} = \bar{V}_{cm} - \bar{V}_{mm} = \partial \bar{V} \quad (4.1) \]

Thus, \( \partial \bar{V}^{(m)} \) depends only on \( \Lambda \). [We shall use \( \partial \) as a dual-wavelength difference operator:

\( \partial( ) = ( )_{cm} - ( )_{mm} \).] The slope \( \Lambda \) and the median volume diameter, \( D_0 \) are related by:

\[ \Lambda = 3.67/D_0 \quad (4.2). \]

The above relation holds for RSD extending from zero to infinite diameter. For a truncated RSD, we regard \( D_0 \) as a proxy for \( \Lambda \). Then, we have:

\[ D_0 = f_v(\partial \bar{V}) \quad (4.3) \]

where the function \( f_v \) depends only on \( \partial \bar{V} \). Thus \( \partial \bar{V}^{(m)} (= \partial \bar{V}) \) is sufficient for retrieving \( D_0 \) and \( \Lambda \).

We have obtained \( f_v \) as follows. First, we calculated the back-scattering and extinction cross-sections for the two wavelengths, as a function of the drop diameter, using methods described in Mishchenko (2003). Calculations were done for horizontally oriented ellipsoidal drops. The drop shape was assumed as given by Pruppacher and Pitter (1971). The \( D_{\text{min}} \) and \( D_{\text{max}} \) were taken as 0.01 and 0.70 cm, respectively. Since the refractive index of water depends on the temperature, the calculations were done for 3 temperatures, namely, 5, 15, and 25 C, covering the range occurring in the observations. The refractive indices used in our calculations are given in Table 2. For the terminal fall speed, we used the analytical approximation given by Atlas et al. (1973) for an air density, \( \rho_0 \), near the ground. The terminal fall speed at a level having air density \( \rho \) is obtained by multiplying the ground-level value by \( (\rho_0/\rho)^{0.4} \) (Foote and
duToit 1969). In practice, we reduced the observed Doppler velocities to the ground level by dividing it by the density factor. Finally, the integrals in equations (2.9) were evaluated numerically as a function of $\Lambda$.

Figure 4a (4b and 4c will be discussed later) shows plots of $\bar{V}_{cm}$, $\bar{V}_{mm}$ and $\partial \bar{V}$ against $D_0$ for a temperature of 15°C. We see that $\partial \bar{V}$ has a maximum at about $D_0 \approx 1.8 \text{ mm}$. At small $D_0$, $\partial \bar{V}$ is small because the scattering is in the Rayleigh regime at both the wavelengths; at large $D_0$, $\partial \bar{V}$ is small because of the saturation of terminal fall velocity with increasing drop size. Using this figure, we can calculate $D_0$ from observed $\partial \bar{V}^{(m)}$. One potential complication is that $D_0$ is a double-valued function of $\partial \bar{V}$. However, the peak of the $\partial \bar{V}$ curve occurs at about $\Lambda = 20 \text{ cm}^{-1}$. According to Marshall and Palmer (1948), $\Lambda = 20 \text{ cm}^{-1}$ corresponds to a rainfall rate of about 30 mm hr$^{-1}$. In stratiform rain, a rainfall rate exceeding 10 mm hr$^{-1}$ is very unlikely. Moreover, in our case, the mm wavelength radiation cannot penetrate any significant length of rain of this intensity. Therefore, in our case, $D_0$ will almost always be smaller than 1.8 mm and can be uniquely estimated from $\partial \bar{V}^{(m)}$. For practical use, it is efficient to use a polynomial fit between $D_0$ and $\partial \bar{V}$:

$$D_0 = \sum_{i=0}^{i=2} a_i (\partial \bar{V})^i$$  \hspace{1cm} (4.4).

A third degree fit was found to give adequate accuracy. The fit is shown in Fig. 1a and the coefficients of the fit are given in Table 3. In estimating $D_0$, the fit for the temperature closest to the temperature at the observation height was used. Thus, if the RSD obeys (1.1), its $\Lambda$ and $D_0$ can be retrieved from the measured differential mean Doppler velocity alone.
Once $D_0$ has been estimated, $w$ can be retrieved simply as the difference between $\bar{V}_{cm}$ calculated from $D_0$ and the measured $\bar{V}_{cm}^{(m)}$ (equation 2.8). Our method of estimating $w$ is superior to methods that use an analytically derived $\bar{Z} - \bar{V}$ relationship (e.g., Rogers 1964) because such relationships assume a value of $N_0$ whereas our method does not.

(ii) Results

Figure 5 shows the measured $\xi_{cm}^{(m)}$ and $\partial\bar{V}_{cm}^{(m)}$, and the retrieved $D_0$ and $w$ fields. The $D_0$ field has trails similar to those in the fields of reflectivity and mean Doppler velocity. The general correspondence between the trails in the $D_0$ and $\xi_{cm}^{(m)}$ fields is noteworthy since $\xi_{cm}^{(m)}$ is not used in the retrieval of $D_0$. The $w$ field also exhibits a trail structure similar to that in the $\xi_{cm}^{(m)}$ field. The updrafts and downdrafts do not appear to be always correlated with high or low reflectivities. This should not be surprising since $w$ is a dynamic quantity which is affected by a number of factors including buoyancy and pressure perturbation forces. The general correspondence between the trail structures of the observed and retrieved parameters gives us confidence in the retrievals.

b. Retrieval of $N_0$ and $R$

(i) Method

So far we have used only the Doppler velocities. In this subsection, we use $Z_{cm}^{(m)}$ to retrieve the intercept $N_0$. For this purpose, we assume that there is no attenuation at the $cm$ wavelength, that is

$$Z_{cm} = Z_{cm}^{(m)}, \quad \xi_{cm} = \xi_{cm}^{(m)}$$

(4.5)
This assumption is reasonable because the cm wavelength radiation suffers negligible attenuation by ice particles occurring in stratiform clouds (e.g., Battan, 1973). In the rain region, the attenuation is larger but still negligible. According to Battan, the two-way rain attenuation \( \kappa_{cm} \approx 0.3, 0.8, 0.02 \text{ dB km}^{-1} \) for \( R = 10, 3 \) and \( 10 \text{ mm hr}^{-1} \) respectively. Since in stratiform rain, \( R \) is generally less than a few mm hr\(^{-1}\), we can neglect attenuation due to rain. Neglecting attenuation by particles in the melting region is more problematical. For now, we assume this attenuation to be also zero; this assumption will be reconsidered later. Now, from (2.2), (4.5) and (1.1), we have:

\[
Z_{cm}^{(m)} / N_0 = Z_{cm} / N_0 = c_z \int_{D_{\text{min}}}^{D_{\text{max}}} \exp(-\Lambda D) \sigma_{\text{BACK,cm}}(D) dD
\]  

(4.6).

The right hand side of this equation is a function of \( \Lambda \) only. Therefore, this equation can be used to estimate \( N_0 \) from the measured \( Z_{cm}^{(m)} \) and the already retrieved \( \Lambda \). Once \( N_0 \) is known, integral parameters of the RSD such as \( R \) can be easily calculated.

(ii) Results

The retrieved \( N_0 \) and the calculated \( R \) are shown in Figs 6a and 6b. The trail structures in the fields of \( N_0 \) and \( R \) also have periodicity similar to that of the trails in the reflectivity field. Figure 7a shows a histogram of \( N_0 \). The \( N_0 \) values vary over more than two orders of magnitude and are generally less than the Marshall-Palmer value. Waldvogel (1974) reported observations of RSD in stratiform rain with a ground-based electromechanical raindrop spectrometer. His time series of \( N_0 \) (see, e.g. his Fig. 2) also show values of \( N_0 \) that are generally less than the Marshall-Palmer value. Uijlenhoet et al. (2003) reported disdrometer observations of RSD for a squall line. For an exponential fit to the RSD, they found values of
$N_0$, smaller than the M-P value in the transition and stratiform regions of the storm. In particular, in the stratiform portion of the rain, their $N_0$ is generally less than $3 \times 10^{-2} \text{cm}^{-4}$ which is the mode value in our histogram plot of $N_0$. Figure 7b shows a scatter plot of $N_0$ against $E_{5\text{cm}}^{(m)}$. There is wide scatter in the values of $N_0$ but it shows a tendency to increase with $E_{5\text{cm}}^{(m)}$ for $E_{5\text{cm}}^{(m)}$ less than 25 $dBZ$. For higher reflectivities, the trend appears to be reversed. This is discussed further below. Figure 7c shows a plot of $N_0$, $D_0$, $w$ and $E_{5\text{cm}}^{(m)}$ against horizontal distance for four selected heights. At all four heights, peaks and troughs in $N_0$ are strongly correlated with troughs and peaks respectively in $D_0$. Similarly, peaks and troughs in $E_{5\text{cm}}^{(m)}$ are fairly well correlated with peaks and troughs in $D_0$, especially for reflectivities smaller than about 25 $dBZ$.

We do not have in-situ observations to support or refute these correlations. However, time series of $\Lambda(=3.67/D_0)$, $N_0$ and $R$ in Figs. 2, 4 and 6 of Waldvogel (1974) also show strong correlations similar to those in our plots. A similar correlation is evident in the observations of Uijlenhoet et al. (see their Fig. 5), especially in the transition region of the precipitation. Such agreement with ground-based observations supports our retrievals. The similarity of correlations in the ground-observed and radar-retrieved parameters is remarkable considering the great disparity in the sample sizes in the two cases. Peaks and troughs in $D_0$ and $w$ also generally occur together although the correlation is not as tight as in the cases of $N_0$ and $D_0$. Figure 7d shows a scatter plot of $E_{5\text{cm}}^{(m)}$ vs. $R$ together with regression equations reported in the literature and a best fit to the data. There is fair agreement between the slopes of the various regression equations. Our best-fit regression is higher than all the other regressions. We see that the increase of $R$ with $Z$ tends to slow down at about 25 $dBZ$. The scatter-plot of $N_0$ vs. $Z$ (Fig. 7b) showed a similar change in trend at about 25 $dBZ$. One possible reason for this change in trends
in Figs. 7b and 7d may be the neglect of attenuation at the cm wavelength. It is possible that melting band attenuation become non-negligible at about 25 dBZ. In the next subsection, we shall see that the retrieved water vapor attenuation also supports this view.

c. Step 3 – Retrieval of Water Vapor Attenuation at the mm Wavelength

(i) Method

So far we have used the measured Doppler velocities and the reflectivity at the cm wavelength to retrieve $N_0$, $D_0$, $w$ and integral parameters, assuming zero attenuation at the cm wavelength. We shall now use the $\xi^{(m)}_{\nu}$ to retrieve attenuation suffered by the mm wavelength radiation.

The measured reflectivity at the mm wavelength is given by:

$$\xi^{(m)}_{\nu}(r) = \xi_{\nu}(r) - \alpha^{(gas)}_{\nu}(0,r) - \alpha^{(ppt)}_{\nu}(0,r_0) - \alpha^{(ppt)}_{\nu}(r_0,r), \quad r \geq r_0 \quad (4.7),$$

where the attenuation is decomposed into that by gases and precipitation. At the mm wavelength, water vapor is responsible for almost all the gaseous absorption. The term $\alpha^{(gas)}_{\nu}(0,r)$ represents the two-way integrated attenuation by water vapor from the radar to the range $r$, and $\alpha^{(ppt)}_{\nu}(0,r_0)$ and $\alpha^{(ppt)}_{\nu}(r_0,r)$ represent, respectively, the attenuations by precipitation particles from the radar to range $r_0$ and from range $r_0$ to the range $r \geq r_0$. We select the range $r_0 \approx 3.8 km$ below the melting region where the ice particles have just melted into raindrops. Then, $\alpha^{(ppt)}_{\nu}(0,r_0)$ stands for attenuation by particles in the ice and the melting regions. Applying (4.7) to range $r_0$:
Thus, the difference between the true and measured reflectivities at range \( r_0 \) is equal to the total integrated attenuation, at the \( mm \) wavelength, from the top of the cloud to range \( r_0 \) by water vapor and precipitation. The attenuation by water vapor at the colder temperatures at ranges less than \( r_0 \) is relatively small; for a saturated atmosphere, the two-way integrated attenuation from the top of the atmosphere down to 4 \( km \) height is about 1.5 \( dB \) (see Fig. 8c below). If we neglect the attenuation by the ‘dry’ ice particles above the melting band, then we can write:

\[
\xi_{\text{mm}}(r_0) - \xi^{(m)}(r_0) = \alpha_{\text{mm}}^{(\text{gas})}(0, r_0) + \alpha_{\text{mm}}^{(\text{ppt})}(0, r_0)
\]  

(4.8).

Thus, the difference between the true and measured reflectivities at range \( r_0 \) is equal to the total integrated attenuation, at the \( mm \) wavelength, from the top of the cloud to range \( r_0 \) by water vapor and precipitation. The attenuation by water vapor at the colder temperatures at ranges less than \( r_0 \) is relatively small; for a saturated atmosphere, the two-way integrated attenuation from the top of the atmosphere down to 4 \( km \) height is about 1.5 \( dB \) (see Fig. 8c below). If we neglect the attenuation by the ‘dry’ ice particles above the melting band, then we can write:

\[
\xi_{\text{mm}}(r_0) - \xi^{(m)}(r_0) = \alpha_{\text{mm,MB}}
\]  

(4.9)

where \( \alpha_{\text{mm,MB}} \) represents the attenuation (\( dB \)), at the \( mm \) wavelength, by the melting band. In general, however, \( \alpha_{\text{mm,MB}} \) should be interpreted as the total attenuation from the top of the cloud to the range \( r_0 \).

To use equations (4.7) – (4.9), we need the true reflectivity, \( \xi_{\text{mm}} \). The true reflectivity can be calculated from the retrieved \( N_0 \) and \( \Lambda \). However, we calculate it without making explicit use of the retrieved \( N_0 \). (This will facilitate the discussion of Retrieval Method II in the next section.) From (4.7) and (4.8), we have:

\[
\left[ \xi_{\text{mm}}(r) - \xi^{(m)}(r) \right] - \left[ \xi_{\text{mm}}(r_0) - \xi^{(m)}(r_0) \right] = \alpha_{\text{mm}}^{(\text{gas})}(r_0, r) + \alpha_{\text{mm}}^{(\text{ppt})}(r_0, r), \ r \geq r_0
\]  

(4.10).

Thus, the difference between the true and measured reflectivities at \( r \) and \( r_0 \) is equal to the sum of the integrated attenuations, between the two ranges, by gases and precipitation. We solve (4.10) for the gas attenuation:
\[ \alpha_{nm}^{(pq)}(r_0, r) = \left[ \xi_{mn}(r_0) - \xi_{mm}(r) \right] - \left\{ \xi_{mm}(r_0) - \xi_{mm}(r) \right\} - \alpha_{nm}^{(pq)}(r_0, r), \ r \geq r_0 \quad (4.11) \]

The term in the square brackets is observed. Again, the term in the curly brackets and the quantity \( \alpha_{nm}^{(pq)}(r_0, r) \) can be calculated using the retrieved \( N_0 \) and \( \Lambda \). Again, we calculate them also without making explicit use of the retrieved \( N_0 \). We write the true reflectivity as:

\[ \xi_{mn} = \xi_{cm} - (\xi_{cm} - \xi_{mm}) = \xi_{cm} - \partial \xi \quad (4.12) \]

The term \( \partial \xi \) is independent of \( N_0 \) and is a function of \( \Lambda \) only. Further, since \( \Lambda \) is a function of \( \partial \bar{V} \) alone, we can express \( \partial \xi \) also as a function of \( \partial \bar{V} \) alone:

\[ \partial \xi = f_\xi(\partial \bar{V}) \quad (4.13) \]

We have evaluated this function numerically (Fig. 1b). In practice, it is efficient to use a polynomial fit:

\[ \partial \xi = \sum_{i=0}^{i=n} b_i (\partial \bar{V})^i \quad (4.14) \]

to calculate \( \partial \xi \) from the observed \( \partial \bar{V}^{(m)} = \partial \bar{V} \). This gives us the terms in the curly brackets of (4.11). The remaining term of (4.11) is evaluated as follows. We have:

\[ \alpha_{nm}^{(pq)}(r_0, r) = c_k \int_{r_0}^r \kappa_{mn}(r') \, dr' \quad (4.15) \]

Now from (1.1), (2.2) and (2.3):

\[ \kappa_{mn} = \frac{c_{k,mn} \int_{D_{mm}}^{D_{max}} \exp(-\Lambda D) \sigma_{EXT,mn}(D) \, dD}{c_{Z,cm} \int_{D_{mm}}^{D_{max}} \exp(-\Lambda D) \sigma_{BACK,cm}(D) \, dD} \quad (4.16) \]
Thus $\kappa_{mn}/Z_{cm}$ is independent of $N_0$ and is a function of $\Lambda$ only. Therefore, it too can be expressed as a function of $\partial \bar{V}$:

$$\kappa_{mn}/Z_{cm} = f_{k/z}(\partial \bar{V}) \quad (4.17).$$

We again fit a polynomial to the function $f_{k/z}(\partial \bar{V})$:

$$\kappa_{mn}/Z_{cm} = f_{k/z}(\partial \bar{V}) = \sum_{i=1}^{i=n} h_i (\partial \bar{V})^i \quad (4.18).$$

Figure 1c shows a plot of $f_{k/z}(\partial \bar{V})$ together with the polynomial fit. The coefficients of the fits are given in Table 3. Using (4.15) and (4.18), we have:

$$\alpha_{mn}^{(\text{ppt})}(r_0, r) = c_k \int_0^r Z_{cm}(r') \left[ \sum_{i=1}^{i=n} h_i \left[ \partial \bar{V}(r') \right]^i \right] dr' \quad (4.19).$$

Thus, $\alpha_{mn}^{(\text{ppt})}(r_0, r)$ can be calculated from the measured $Z_{cm}^{(m)} = Z_{cm}$ and $\partial \bar{V}^{(m)} (= \partial \bar{V})$. Equation (4.11) then yields the integrated water vapor attenuation $\alpha_{mn}^{\text{wat}}(r_0, r)$. An estimate of specific attenuation by water vapor can be obtained by differentiating $\alpha_{mn}^{\text{wat}}(r_0, r)$. The specific attenuation can be translated into water vapor content using a temperature sounding and the known dependence of the specific attenuation on temperature, pressure and water vapor content (Ulaby et al. 1982).

(ii) Results

Figure 8a shows $\alpha_{mn, MB}$, the retrieved two-way attenuation at the $mm$ wavelength by precipitation particles and the water vapor from the top of the cloud to just below the melting band, and $\xi_{cm, MB}$, the maximum $cm$-wavelength reflectivity in the melting band, as functions of the horizontal distance. Except for a small noisy region near 126 km, all the retrieved
attenuations are positive. Moreover, there is a strong correlation between the peaks and troughs of the two quantities lending further support to the retrieval.

The retrieved two-way integrated water vapor attenuation, $\alpha_{\text{mm}}^{\text{gas}}(r_0, r)$, between 135 & 170 km is shown in the Fig. 8b (the mm-wave signal is noisy in the other regions). First, we notice that the attenuation is highly variable. This means that the water vapor distribution must also be highly variable. We also see a trail structure in the integrated attenuation. This should not be surprising since the water vapor is affected by evaporation / condensation, which depends on the RSD and the vertical air velocity and these latter quantities have trail structures. Secondly, the total integrated attenuation occasionally exceeds that calculated for a saturated atmosphere (7.2 dB). As argued later, this unphysical result may be due to our assumption of zero attenuation at the cm wavelength. Thirdly, the integrated attenuation sometimes decreases with distance downwards. This is also an unphysical result because the integrated attenuation should increase downwards monotonically. This is better seen in the vertical profiles for four selected horizontal distances (Fig. 9).

For each distance, Fig. 9 shows three sets of measured and retrieved quantities, namely, (left) $\xi^{(m)}_{\text{cm}}, \xi^{(m)}_{\text{mm}}, \xi_{\text{mm}}$, (middle) $\overline{V}^{(m)}_{\text{cm}}, \overline{V}^{(m)}_{\text{mm}}$, and (right) $\alpha^{(\text{gas})}_{\text{mm}}, \alpha^{(\text{ppt})}_{\text{mm}}$ and integrated gas attenuation computed for a saturated atmosphere from the sounding using equations in Ulaby et al. (1982). The following discussion applies mostly to the right panels.

At 138.5 km, the retrieved integrated water vapor attenuation $\alpha^{(\text{gas})}_{\text{mm}}$ increases downwards monotonically except for small decreases at certain heights. These decreases may be due to signal fluctuations. Comparison of retrieved and computed attenuations suggests an unsaturated atmosphere with a relative humidity of about 50%. The integrated rain attenuation below $r_0$ is
about 6.5 dB. At 142 km also, $\alpha_{nn}^{(gas)}$ increases downward almost monotonically but there is a rather thick layer of supersaturated air below about 1.5 km height. Again at 144 km, $\alpha_{nn}^{(gas)}$ has an almost monotonic profile except for a rather deep downward decreasing section just below $r_0$. This profile also has an unphysical shallow supersaturated column below about 0.5 km. The integrated rain attenuation below $r_0$ is about 4 dB. The supersaturated column may be due to an underestimate of attenuation by precipitation, which in turn may be due to the neglect of attenuation at the cm wavelength as discussed later. The profile of $\alpha_{nn}^{(gas)}$ at 152 km is the most anomalous. This anomaly is also quite evident in Fig. 8b. It probably involves a breakdown of our assumption of an exponential RSD. Because of the occasional non-monotonic nature of the retrieved integrated attenuation, we have not attempted to recover the distribution of water vapor.

5. Retrieval II: Method and Results

The first retrieval method gave us some physically meaningful results but we also noted certain deficiencies in the results: (a) the integrated water vapor attenuation occasionally decreased with distance downwards, (b) occasionally the water vapor attenuation implied supersaturated air, and (c) the plots of $S_{cm}^{(m)}$ vs. $N_0$ and $S_{cm}^{(m)}$ vs. $R$ showed a change in trend at a reflectivity of about 25 dBZ. As regards (a), decreases of small magnitudes over a few range bins may be due to signal fluctuations but decreases of larger magnitudes over many range bins are problematic. The more significant anomalies may involve breakdown of the assumption of exponential RSD. Discrepancies (b) and (c) may be due, at least in part, to neglect of attenuation at the cm-wavelength. As pointed out earlier, neglect of attenuation by raindrops and ice particles is justifiable. However, attenuation by mixed phase and melting particles may start to become significant at about 25 dBZ. Bellon et al. (1997) reported melting band attenuation at 3.2 cm-wavelength of about 0.5 dB (1.7 dB) for a rain reflectivity, below the melting band,
of 23 dBZ (36.5 dBZ). The Bellon et al. measurements were made in temperate-latitude precipitation, whereas the observations discussed here were made in tropical precipitation. For a given reflectivity, the melting band attenuation could be greater in tropical precipitation because larger, more deformed, mixed-phase particles which are responsible for the attenuation are more likely in tropical precipitation. Qualitatively, we can see the effect of non-zero melting band attenuation on the retrievals of the previous section. First, we note that the retrieved $A$, $D_0$, and $w$ will remain unchanged because those retrievals are unaffected by attenuation. Second, with finite melting-band attenuation at the $cm$-wavelength, $Z_{cm}$ would exceed $Z_{cm}^{(m)}$ for $r \geq r_0$. Consequently, $N_0$ retrieved assuming non-zero melting band attenuation would be greater than that deduced by Method I. Thus, with finite melting band attenuation, the retrieved RSD would have the same slope but a greater intercept than that retrieved by Method I. Because of the increased intercept, the estimated attenuation by precipitation would be greater, which would result in reduced values of the retrieved integrated water vapor attenuation. This may help to resolve problem (b) above. Also, it may be shown that neglect of melting band attenuation tends to produce the tendencies noted in (c) above.

Below, we present a method for retrieving the melting band attenuation at the $cm$-wavelength from the measurements. Results of application of this method are presented in subsection b. A synopsis and comparison of the results from methods I and II are presented in subsection c.

a. Method

To retrieve the melting band attenuation at the $cm$-wavelength, we need the profile of water vapor attenuation at the $mm$-wavelength; it can be calculated if the distribution of water vapor is known. For this purpose, we use the temperature sounding of Fig. 2 and assume a saturated atmosphere. However, the method can be applied to any given distribution of water vapor.
The first step is the retrieval of \( A, D_0 \) and \( w \) as in section 4a (i). Next, denoting the melting band attenuation at the \( cm \) – wavelength by \( A_{cm,MB} \) (\( \alpha_{cm,MB} \) dB), equation (4.5) gives:

\[
Z_{cm} = Z_{cm}^{(m)} / A_{cm,MB}, \quad \xi_{cm} = \xi_{cm}^{(m)} + \alpha_{cm,MB}, \quad r > r_0
\]  

(5.1)

Thus, for \( r \geq r_0 \), the true reflectivity (dBZ) at the \( cm \) – wavelength exceeds the measured reflectivity (dBZ) by a constant equal to the melting band attenuation in dB. The intercept of the RSD is given by the following equation (see equation 4.6):

\[
N_0 = \left[ Z_{cm}^{(m)} / A_{cm,MB} \right] \int_{D_{\min}}^{D_{\max}} \exp(-\Lambda D) \sigma_{BACK,cm} (D) dD = \frac{N'_0}{A_{cm,MB}}
\]  

(5.2)

where \( N'_0 \) is the intercept under the assumption of zero melting band attenuation at the \( cm \) – wavelength. We see that \( N_0 > N'_0 \).

The second step involves estimation of \( A_{cm,MB} \). For this, we use the measured reflectivities at both wavelengths. From (4.7) and (5.1), we have:

\[
\xi_{cm}^{(m)} (r) - \left[ \xi_{mn}^{(m)} (r) + \alpha^{(gas)}_{mn} (0,r) \right] - \left[ \xi_{cm} (r) - \xi_{mn} (r) \right] = \alpha^{(ppr)}_{mn} (r_0, r) + \left[ \alpha^{(ppr)}_{mn} (0,r_0) - \alpha_{cm,MB} \right], \quad r \geq r_0
\]  

(5.3)

or

\[
\partial \xi^{(m)}_p (r) - \partial \xi (r) = \left\{ \alpha^{(ppr)}_{mn} (r_0, r) - \partial \alpha_{MB} \right\}, \quad r \geq r_0
\]  

(5.4)

where

\[
\partial \xi^{(m)}_p (r) \equiv \xi^{(m)}_p (r) - \left[ \xi^{(m)}_p (r) + \alpha^{(gas)}_{mn} (0,r) \right]
\]  

(5.5)

\[
\partial \xi (r) \equiv \left[ \xi_{cm} (r) - \xi_{mn} (r) \right]
\]  

(5.6)

and

\[
\partial \alpha_{MB} \equiv \alpha_{cm,MB} - \alpha^{(ppr)}_{mn} (0,r_0) \equiv \alpha_{cm,MB} - \alpha_{mn,MB}
\]  

(5.7)
The term, in square brackets on the left hand side of (5.3) is the \textit{mm} wavelength reflectivity that would be observed in the absence of gas attenuation. Together, the first two terms on the left hand side thus represent the difference of the reflectivities at the two wavelengths due to precipitation alone. This quantity is denoted by the symbol $\partial \xi^{(m)}_p$ (Eq. 5.5). The last term on the left hand side of (5.3) is the difference of the true reflectivities at the two wavelengths and is denoted by $\partial \xi$ (Eq. 5.6); in the literature, it is often called the dual-wavelength ratio. The left hand side of (5.3) is thus the difference between the measured and true differential reflectivities.

Then the right hand sides of (5.3) (and 5.4) are equal to the difference between the integrated attenuations from the radar to the range under consideration at the \textit{mm} and \textit{cm} wavelengths. This differential attenuation is split into attenuation between the radar and the range $r_0$ and between $r_0$ and $r (\geq r_0)$. The latter quantity is simply the rain attenuation at the \textit{mm} wavelength between $r_0$ and $r$ since we assume rain attenuation at the longer wavelength to be zero. At the longer wavelength, the only attenuation between the radar and $r_0$ is due to the melting band. If we assume that at the shorter wavelength also the attenuation is due to the melting band only then the former quantity is simply the differential melting band attenuation, equation (5.7). A special case of equation (5.3) occurs when the scattering at both wavelengths is in the Rayleigh regime and there is no attenuation at the longer wavelength. Then $\partial \xi$ equals zero, and the term in the curly brackets, in equation (5.4), is simply the integrated attenuation by precipitation at the shorter wavelength. These conditions are fairly well satisfied by radiation of wavelengths 10 and 3 \textit{cm} (e.g. Eccles and Mueller 1971).

Now we consider application of equation (5.4) to the observations. The first term of this equation is given by observation and the assumed water vapor distribution. The second term can be evaluated from the observed $\partial \bar{V}$ using equation (4.13). This gives an estimate of the term in
the curly brackets in (5.4). Putting \( r = r_0 \) in this equation, we obtain an estimate of the
differential melting band attenuation:

\[
\partial \alpha_{MB} = \alpha_{cm,MB} - \alpha_{mm,MB} = \partial \xi(r_0) - \partial \xi^{(m)}(r_0)
\]  

(5.8).

Once \( \partial \alpha_{MB} \) is known, we can apply (5.4) to ranges \( r > r_0 \) to estimate the integrated attenuation
by precipitation:

\[
\alpha_{mm}^{(ppr)}(r_0, r) = \partial \alpha_{MB} + \partial \xi^{(m)}(r) - \partial \xi(r) \\
= [\partial \xi^{(m)}(r) - \partial \xi^{(m)}(r_0)] - [\partial \xi(r) - \partial \xi(r_0)]
\]

(5.9).

It is noted that the above estimate of \( \alpha_{mm}^{(ppr)}(r_0, r) \) involves the measured differential Doppler
velocity, the measured reflectivities at both wavelengths and the assumed water vapor
distribution. We can get a second estimate of \( \alpha_{mm}^{(ppr)}(r_0, r) \) using only the measured differential
Doppler velocity and the measured reflectivity at the cm-wavelength. Using (4.15), (4.17), and
(5.1), we have:

\[
\alpha_{mm}^{(ppr)}(r_0, r) = c_s \int_0^r \kappa_{mm}^{(m)}(r') \, dr' = \frac{c_s}{A_{cm,MB}} \int_0^r \frac{\kappa_{mm}^{(m)}(r')}{Z_{cm}(r')} Z_{cm}^{(m)}(r') \, dr' \\
= \frac{c_s}{A_{cm,MB}} \int_0^r f_{\gamma} \left( \partial \tilde{V}(r') \right) Z_{cm}^{(m)}(r') \, dr' = \frac{\alpha_{mm}^{(ppr)}(r_0, r)}{A_{cm,MB}}
\]

(5.10).

In the above equation, \( \alpha_{mm}^{(ppr)}(r_0, r) \) represents the integrated attenuation by precipitation under
the assumption of zero attenuation at the cm-wavelength. Equations (5.9) and (5.10) give two
independent estimates of \( \alpha_{mm}^{(ppr)}(r_0, r) \) which should be identical. This condition can be used to
estimate \( A_{cm,MB} \) and then \( \alpha_{mm,MB} \) can be retrieved from \( \partial \alpha_{MB} \) (Eq. 5.7). A method for retrieving
these quantities is illustrated in the schematic of Fig. 10. The curve marked A represents \( \partial \xi^{(m)}_{zp} \).
Since it involves the observed reflectivities, it is noisy. The curve marked B represents \( \partial \xi \); it is
calculated from the observed \( \partial \tilde{V}^{(m)} \) and, as such, it is also noisy. The difference of these two
curves is labeled C and equals $\alpha_{mn}^{(pp)}(r_0,r) - \partial\alpha_{MB}$. At the height marked $r_0$, the displacement of this curve from the ordinate should equal $-\partial\alpha_{MB}$. However, we do not take this displacement as an estimate of $-\partial\alpha_{MB}$ because of the noisy nature of the curves. The curve labeled D represents $\alpha_{mn}^{(pp)}(r_0,r)$ and is obtained by evaluating the integral in the fourth member of equation (5.10) numerically. This curve turns out to be relatively smooth because it is the integral of a positive quantity. In practice, it is also found to be fairly linear. The curve D when divided by $A_{cm,MB} < 1)$ and displaced to the right by the amount $-\partial\alpha_{MB}$ should coincide with the curve C. Fitting techniques can be used to find the displacement and $A_{cm,MB}$ to achieve the best fit between curves D and C. However, in view of the rather questionable validity of the assumption of a saturated atmosphere, we use a simple linear least squares fitting technique. First, we smooth the curve C by making a linear least squares fit to it. The line labeled L shows the fitted line. A few range gates nearest to the ground are excluded from this fitting procedure because of possible ground contamination of those observations. The displacement of the fitted line from the ordinate at range $r_0$ is taken as an estimate of $-\partial\alpha_{MB}$ and the ratio of the slopes of the lines D and L gives an estimate of $A_{cm,MB}$.

b. Results from Retrieval Method II

Because of uncertainty regarding the distribution of water vapor, the following results should not be regarded as definitive but rather as illustrative of the retrieval method; our main purpose here is to develop a retrieval method that can be used if the water vapor distribution is known or measured independently.

Figures 11 shows profiles of the quantities in the schematic of Fig. 10 for the same four horizontal distances as in Fig. 9. The melting band attenuations at the mm wavelength (strictly, attenuation by gases and precipitation from the top of the cloud to the range $r_0$ obtained by
methods I and II are comparable. For method I, the values are 7.80, 4.18, 5.42, 6.42 dB while for method II they are 6.89, 5.54, 5.62, 5.40 dB. The retrieved melting band attenuations for the cm wavelength are: -0.46, 3.79, 3.07, -4.91 dB. The negative attenuations are unphysical and the positive attenuations are higher than would be expected on the basis of Bellon et al.'s measurements. These discrepancies bring out the importance of the water vapor attenuation profile for the retrieval method. The ratio of the slopes of the best-fit line L and the line D gives $1/A_{cm,MB}$. For a positive melting band attenuation, this ratio should be greater than 1. In Fig. 11a, it is slightly less than 1, which yields a negative $\alpha_{cm,MB}$. The slope of the line L is determined by line C, which is obtained from the observed reflectivities and the assumed integrated gas attenuation. If we had assumed smaller gas attenuation, then the line C would be rotated counterclockwise and would have a greater slope, which would then have yielded a positive value for the melting band attenuation. In support of this, we see from Fig. 9a that Method I, which assumed zero melting band attenuation at the longer wavelength, gave an integrated gas attenuation about half that for a saturated atmosphere. In Figs. 11b and 11c the ratio of the slopes mentioned above is greater than unity and consequently the melting band attenuations are positive but rather large in magnitude. Again, an adjustment of the assumed water vapor distribution could alter these results in a more plausible direction. Figure 11d yields a result similar to that of Fig. 11a and is again probably due to an overestimate of water vapor attenuation.

c. Horizontally Averaged Results from Retrieval Methods I and II

In this subsection, we present profiles of horizontal averages of certain observed and retrieved quantities for the region 135–155 km. Only the region of 135–155 km were used in the averaging because the received power, at the shorter wavelength, is near or below the noise
level in the other regions. Profiles that yielded negative melting band attenuation at the longer wavelength have also been ignored in this averaging.

Figure 12a shows the measured reflectivities for the two wavelengths and the retrieved ‘true’ reflectivity at the longer wavelength. The strong attenuation suffered by the $\mu m$ wavelength is obvious. The measured and ‘true’ reflectivities, at the $cm$ wavelength, differ by a constant amount equal to the average melting band attenuation for this wavelength deduced by Method II. This difference, about $2.5 \, dB$, is rather large and is probably overestimated because of the assumption of a saturated atmosphere in Method II. The downward decrease of the reflectivity at the $cm$ wavelength is probably due to evaporation and drop breakup. The total decrease of reflectivity from the top to the bottom of the column is about $5 \, dB$.

Figure 12b shows measured mean Doppler velocities and the retrieved vertical air velocity. As mentioned earlier, the retrieved $w$ is the same for both methods. There is an updraft above about $2.7 \, km$ and downdraft below. The peak upward velocity of about $25 \, cm \, s^{-1}$ occurs at the top of the column; the downdraft increases downwards reaching a maximum magnitude of about $25 \, cm \, s^{-1}$ at about $1 \, km$ above ground. Below $1 \, km$, the downdraft shows a satisfactory decrease towards the ground. The pattern and magnitudes of the retrieved vertical air velocities are characteristic of extensive stratiform rain (e.g., Houze, 1993, Chapter 9; Srivastava et al, 1986). The studies in the literature have used VAD or other methods to infer area-averaged vertical air velocities. Here we have been able to deduce vertical velocities in stratiform rain from vertical incidence Doppler radar observations at two wavelengths.

Figure 12c shows the median volume diameter, $D_o$, and the intercept parameter, $N_0$. The $D_o$ is the same for both retrieval methods. The $N_0$ is greater for Method II. It increases rather rapidly from the top of the column to a height of about $2 \, km$ and is approximately
constant below that height. The $D_0$ also decreases rapidly down to a height of about 2 km and more slowly below that height. The rainfall rate deduced by both methods (Fig. 12d) is approximately constant from the top of the column down to a height of about 2 km but decreases below that height. The rainwater content (Fig. 12d) shows a small increase (Method I) from the top of the column to a height of about 2 km and decreases below that height. The decreases in $N_0$ and $D_0$ from the top of the column to about 2 km height are consistent with breakup of drops formed by the melting of large aggregate snowflakes. This is further supported by the near constancy of the rainfall rate in this height interval. For a fixed $R$, the decrease in $D_0$ implies an increase in the rainwater content. The constancy of $N_0$ and the slower decrease of $D_0$ below 2 km height are consistent with rain evaporation. This is also supported by the downward decrease of rainfall rate in this region. We also note that there is downdraft in the region where the rainfall rate and rainwater content are decreasing with distance downwards. The above deductions are physically reasonable and consistent but should be regarded as somewhat speculative because of the lack of independent supporting measurements. In future, additional measurements, perhaps at a third frequency, could be used to refine and extend the retrievals and further support the physical deductions.

Figure 12e shows profiles of integrated water vapor attenuation retrieved by Method I and that calculated for a saturated atmosphere. The averaged attenuation increases monotonically downwards and shows a nearly saturated or a somewhat supersaturated atmosphere. Individual profiles presented earlier sometimes showed sections having an unphysical decrease downwards or rather large supersaturations. These are not seen in the averaged profile. It is possible that the averaging has served to remove errors due to signal fluctuations and beam mismatch. The small supersaturation in the lower levels is probably due to the neglect of melting band attenuation in
Method I. Another possibility is that the sounding used to compute the attenuation may not be quite applicable to the region of the radar measurements.

6. Summary and Discussion

We have presented observations of light stratiform rain with melting band by airborne dual-wavelength, nadir-looking radars. The radars measure reflectivities and mean Doppler velocities at two wavelengths - 3.2 cm and 3.2 mm. Except for very large raindrops, the scattering at the longer wavelength is in the Rayleigh regime; at the shorter wavelength it is in the Mie regime except for the very smallest raindrops. There is also strong attenuation at the shorter wavelength. Consequently, the mean Doppler velocities and reflectivities measured at the two wavelengths are different. These differences have been exploited to retrieve RSD parameters, water vapor and precipitation attenuation, melting band attenuation and vertical air velocity. Two retrieval methods have been presented. In both methods, we assume the RSD to be exponential with two free parameters - slope, \( \Lambda \), and intercept, \( N_0 \). In both methods, the \( \Lambda \) (equivalently \( D_0 \)) and the vertical air velocity have been estimated from the mean Doppler velocities alone. In Method I, the \( N_0 \) has been retrieved from the measured reflectivity at the longer wavelength assuming no attenuation at this wavelength. From the retrieved \( \Lambda \) and \( N_0 \), integral parameters of the RSD such as rainwater content and rainfall rate have been calculated. Additionally, in Method I, the measured reflectivity at the shorter wavelength has been used to retrieve attenuation at this wavelength by the melting band and the water vapor below it. We believe that this is the first time that range-resolved attenuation by water vapor has been retrieved by active radar in a precipitating situation. In retrieval method II, we dropped the assumption of zero attenuation at the longer wavelength to the extent of assuming a finite melting band attenuation. Under the further assumption of a known distribution of water vapor, we have presented a method for retrieving the melting band attenuations at both wavelengths and
the rain attenuation at the shorter wavelength. For illustration, Method II has been applied assuming a saturated atmosphere.

Lacking independent in-situ and/or ground based measurements, we have argued in favor of the validity of our retrievals on the basis of physical plausibility and consistency. The retrieved parameters such as $D_0$, $\Lambda$, and the vertical air velocity have trail structures similar to those of observed reflectivity and mean Doppler velocity at the longer wavelength. A strong negative correlation found between the peaks and troughs of retrieved $D_0$ and $N_0$ is similar to that reported in ground-based observations in stratiform precipitation by others. The numerical values and scatter of the retrieved $N_0$ are also similar to those found in ground-based studies of stratiform rain RSD. The retrieved melting band attenuation at the shorter wavelength has shown strong positive correlation with melting band reflectivity at the longer wavelength. The profiles of the integrated water vapor attenuation retrieved by Method I have been found to be generally increasing with distance downwards as would be expected. The profiles generally implied a subsaturated atmosphere as is physically plausible for extensive stratiform rain. Small increases of the retrieved integrated water vapor attenuation with distance downwards have been attributed to signal fluctuations, while larger anomalies have been attributed to breakdown of the assumption of exponential RSD. Some supersaturated regions have also been indicated by the retrievals, especially in the lower levels. These have been attributed to the neglect of attenuation at the longer wavelength. Accordingly, in retrieval Method II, we dropped the assumption of zero melting band attenuation at the longer wavelength. Application of Method II requires knowledge of the distribution of water vapor. The illustrative retrievals in section 5 assumed a saturated atmosphere. Application of Method II yielded $D_0$ and $w$ fields identical to those given by Method I. For Method II, the $N_0$ has been found to be greater than that for Method I. The retrieved melting band attenuation at the shorter wavelength was similar to that for Method I.
The retrieval occasionally yielded unphysical negative melting band attenuations at the longer wavelength. This has been attributed to an overestimate of the water vapor attenuation by our assumption of a saturated atmosphere. Summary results have been presented in the form of profiles of horizontal averages of observed and retrieved quantities. The averaged vertical air velocity is noteworthy. The magnitudes and vertical variation of the velocity are remarkably similar to what has been reported in the literature for area-averaged vertical velocity in extensive stratiform rain measured by the VAD method. The averaged profile of rainfall rate, $N_0$ and $D_0$ has shown evidence of drop breakup below the melting band and evaporation in the lower levels in the region of downdrafts. The potential ability to retrieve the parameters of the RSD, the vertical air velocity, and the water vapor distribution as functions of the height is exciting because it offers the possibility of direct calculation of profiles of evaporation and, therefore, latent cooling in the atmosphere using microphysical principles.

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Appendix A
List of Symbols

\( A(\alpha) \): Attenuation factor, number \((dB)\)

\( a_i, b_i, h_i \): Fit coefficients for \( D_0, \partial \xi, \) and \( k_{\text{mm}}/Z_{\text{cm}} \), \( i = 0, 1, 2, 3 \)

\( c_i(\cdot) \): Unit Conversion Factor for Quantity \((\cdot)\)

\( D_0 \): Medium Volume Diameter

\( k(\kappa) \): Two-Way Specific Attenuation / Length, \((dB\ km^{-1})\)

\( N_0 \): Intercept of Raindrop Size Distribution

\( N(D) \): Raindrop Size Distribution

\( V_t \): Terminal Fall Speed

\( \bar{V} \): Mean Doppler Velocity

\( w \): Vertical Air Velocity

\( Z(\xi) \): Reflectivity, \( mm^6 m^{-3} (dBZ) \)

\( \partial (\cdot) = (\cdot)_{\text{cm}} - (\cdot)_{\text{mm}} \)

\( \Lambda \): Slope Parameter

\( \sigma \): Cross-section

Subscripts

\( BACK \): Back-Scattering

\( cm \): 3.2 \( cm \) Wavelength

\( EXT \): Extinction

\( MB \): Melting Band

\( mm \): 3.2 \( mm \) Wavelength

\( p \): Precipitation

Superscripts

\( (\text{gas}) \): Gas, Water Vapor

\( (m) \): Measured

\( (ppt) \): Precipitation
Reference


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Table 1. Characteristics of the radars

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Table 2. Refractive Index for temperatures of 5, 15, 25 C.

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Table 3. Fitting coefficient of polynomials for $D_0$, $\xi_{\text{cm}} - \xi_{\text{mm}}$, and $k_{\text{mm}} IZ_{\text{cm}}$, $0.01 \text{ mm} < D_0 < 1.18 \text{ mm}$.

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