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Dynamical, microphysical and radiative properties of ice clouds using Doppler cloud radar measurements

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1 Introduction

The knowledge of the cloud properties has been recently identified as a mandatory step to reach if the operational weather and climate change forecasts are to be improved. In the framework of the future space missions devoted to the monitoring of the microphysical, radiative, and dynamic properties of clouds at global scale using cloud radar and lidar combination (CLOUDSAT/CALIPSO as part of the Afternoon Train), there is a need for ground-based and airborne validation of the radar/lidar measurements and products from these space missions. The synergy between the two instruments is such that in moderately thick clouds the liquid/ice water content and effective radius of droplets/crystals can be accurately retrieved from these two measurements. The domain of application of the radar-lidar synergy is however limited to a given range of clouds (optical thickness less than 3, roughly). As an example, prefrontal clouds and mixed-phase clouds, which are very common in midlatitude regions, are generally not fully traversed by the lidar. In the present paper we therefore propose an original method complementary to the radar-lidar algorithm, which makes use of the three measurements of a Doppler cloud radar (35 or 95 GHz), namely the radar reflectivity, the Doppler velocity, and the spectral width of the Doppler spectrum in order to recover the effective radius and terminal fall velocity of crystals, the ice water content, and the visible extinction, and therefore the visible optical depth. This radar retrieval method is described in Sect. 2. It relies on a set of statistical relationships between the cloud properties and the radar measurements, scaled by the intercept parameter of the normalized particle size distribution (Testud et al., 2001). These statistical relationships are derived from a large database of airborne in-situ microphysical data collected in both midlatitude and tropical regions. The results obtained with this method and near future work is finally discussed in Sect. 3.

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2 The retrieval of ice cloud properties using a Doppler cloud radar

2.1 The normalized particle size distribution and statistical relationships

Both the three radar observables and the microphysical and radiative parameters that are required to document the cloud properties depend on the particle size distribution (hereafter referred to as PSD) in the radar volume. As a result, any forward model between radar observables and cloud properties includes the statistical properties of the PSD. It is well known that this PSD is highly variable in both liquid and ice phases, owing to variations over three to four decades of the ice water content in a single cloud, and the very different ranges of diameters encountered from one cloud to another. Testud et al. (2001) have recently proposed a formalism that allows a comparison of very different PSDs in the liquid part of precipitating systems. This formalism, known as the normalized PSD, consists in scaling the diameter and concentration axes in such a way that the PSDs become independent of the ice water content IWC and the mean volume-weighted diameter D_m (see Delanoë et al. (2004) for further details). A general expression of the normalized PSD can be written as:

$$N(D_{eq}) = N_0^* F(D_{eq}/D_m) \qquad (1)$$

where $N(D_{eq})$ is the PSD, D_{eq} the equivalent-melted diameter, N_0^* the intercept parameter of the PSD, F the analytical shape of the PSD. The relationship between the true diameter and the equivalent-melted diameter involves an assumption on the ice crystal density, which is a critical point for all methods (Delanoë et al., 2004). The way this assumption is dealt with in the present radar method will be discussed in the next subsection.

Using this formalism, the ice water content can be analytically expressed as a function of N_0^* and D_m through:

$$IWC = \frac{N_0^* \pi \rho_w D_m}{4^4} \qquad (2)$$

Delanoë et al. (2004) have investigated the stability of the PSD shape in ice clouds, using a very extensive airborne in-situ microphysics dataset, including different types of ice clouds, and both midlatitude (CLARE98, CARL99, CARL2000, CARL2001, EUCREX, FASTEX, ARM IOP) and tropical (CEPEX, CRYSTAL-FACE) campaigns. They found that as in the case of the raindrop size distribution the shape of the normalized PSD was fairly stable, and therefore proposed to use a single analytical formulation (the so-called gamma μ shape) to describe ice cloud PSDs:

$$F_\mu(D_{eq}/D_m) = \frac{(4+\mu)^{4+\mu}\Gamma(4)}{4^4\Gamma(4+\mu)}\left(\frac{D_{eq}}{D_m}\right)^{\mu} e^{-(4+\mu)\frac{D_{eq}}{D_m}} \quad (3)$$

This stability in shape is an important result, since it also implies that all moments of the normalized PSD can be related to each other by a power-law relationship. When assuming the gamma μ shape, the ith moment of the PSD can be analytically expressed as:

$$M_i = N_0^* \frac{\Gamma(4)}{\Gamma(4+\mu)} \frac{(4+\mu)^{3-i}}{4^4} \Gamma(i+\mu+1) D_m^{i+1} \quad (4)$$

The effective radius R_e can be defined as the ratio of the third moment to the second, while the mean volume-weighted diameter is the ratio of the fourth to the third moment. This translates into a direct analytical relationship between R_e and D_m:

$$R_e = \frac{\Gamma(7)}{2\Gamma(6)} D_m \quad (5)$$

The general expression of the PSD moments can be used to relate the cloud parameters and radar observables (that depend on different moments of the normalized PSD) through statistical relationships. The radar reflectivity Z (assuming Rayleigh scattering) is proportional to the sixth moment of the PSD, while the visible extinction is proportional to the second moment. It follows from this that there is a direct power-law relationship between Z/N_0^* and α/N_0^*. Therefore, if N_0^* is known and Z is measured, α can be estimated using this relationship (given in Delanoë *et al.* 2004). In Mie regime, this analytical relationship does not hold anymore, but a fit can be performed for the larger ice particles to establish such a relationship between Z/N_0^* and α/N_0^*. The estimate of α can finally be integrated in the vertical to access the cloud optical depth.

In conclusion, if we can estimate the ice density, N_0^*, and D_m, then we can access an extensive documentation of the ice clouds, including ice water content, effective radius, visible extinction, and cloud optical depth. The method proposed in the present paper consists in estimating these quantities from the three radar measurements (radar reflectivity, mean Doppler velocity, and spectral width). This method is described in the next section.

2.2 Ice density estimate from the three radar moments

The first step of the method is to estimate the ice density to be used in the calculations of the previous subsection. In the present method we propose to access such an information from the radar moments.

The reflectivity-weighted terminal fall velocity of a population of ice crystals described by a given PSD can be expressed as:

$$V_T = \frac{\int N(D)v(D)\sigma(D)dD}{\int N(D)\sigma(D)dD} \quad (6)$$

where $v(D) = AD^B$ is the terminal fall velocity of an individual ice particle of diameter D, and $\sigma(D)$ is the radar backscattering cross section. The $v(D)$ relationship is related to the ice density assumption that has to be made in all this study. Matrosov et al. (2002) have proposed to estimate D_m from the terminal fall velocity derived from the mean Doppler velocity measured by a 35 GHz radar. In their method, they have developed a relationship between A and B, which can be written as:

$$B = 0.17A^{0.24} \quad (7)$$

They have also observed a correlation between A and D_m, that can be expressed as another power-law relationship. They conclude that the uncertainty of their procedure to estimate D_m from V_T is around 35%, which is not negligible, although sufficient for many applications. In the present paper we propose an alternative approach, which consists in retrieving the B coefficient of the $v(D)$ relationship from the radar moments and make use of the relationship between B and A derived by Matrosov et al. (2002). It is expected that the retrieval of B will reduce the uncertainty in the Matrosov method arising from the use of an $A-D_m$ relationship. This method consists in computing the ratio of the Doppler spectral width to the terminal fall velocity.

The Doppler spectral width can be written as follows:

$$\sigma_D = \sqrt{\frac{\int N(D)\sigma(D)v^2(D)dD}{\int N(D)\sigma(D)dD} - \overline{V_T}^2} \quad (8)$$

As seen from Eqs. (6) and (8), the A coefficient of the $v(D)$ relationship cancels out in the σ_D/V_T ratio, which leaves a direct analytical relationship between the B coefficient and the σ_D/V_T ratio. This relationship only depends on the analytical shape assumed for the normalized PSD. Figure 1 shows this relationship when the gamma μ analytical shape of Eq. (3) is assumed. It is clearly seen that in this case the σ_D/V_T ratio is a roughly linear function of B. Finally, when B is retrieved from σ_D/V_T, the A coefficient can be estimated using Eq. (7). As discussed in Matrosov et al. (2002) and many others, each crystal habit can be categorized by such a $v(D)$ relationship, so when this relationship is known, a corresponding ice density can be assumed. This is the principle we have adopted here. This part of the method is not automated though, so we still have to choose a single ice density relationship for a given ice cloud, but in a near future

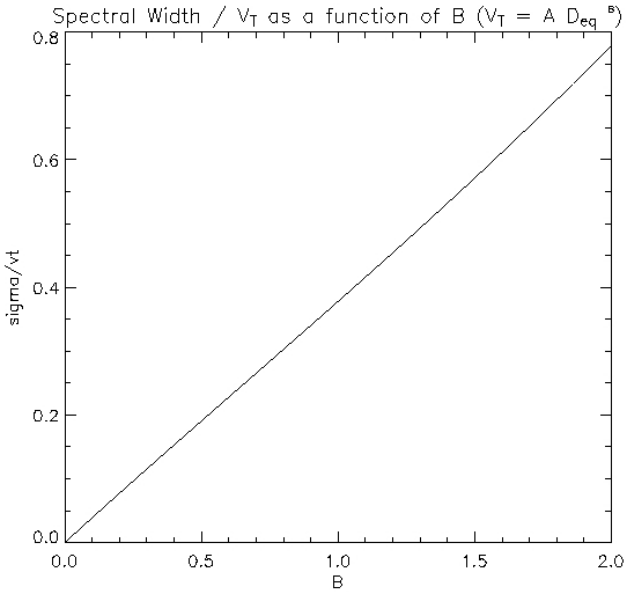

Fig. 1. Theoretical relationship between the ratio of Doppler spectral width to terminal fall velocity and the B coefficient of the $v(D)=AD^B$ relationship.

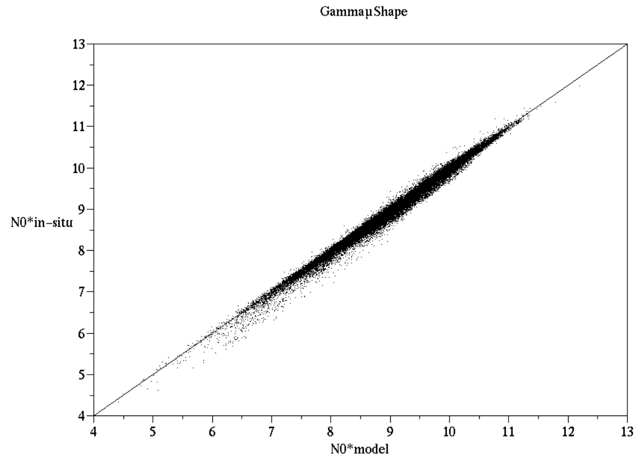

Fig. 2. N_0^* retrieved using the analytical shape of Eq. (3) as a function of the true N_0^* computed from the true PSDs for an extensive airborne in-situ microphysics database.

we will select for each radar gate the best ice density to be applied as a function of the retrieved $v(D)$ relationship.

A Doppler cloud radar does not measure directly the terminal fall velocity. The Doppler measurement is the sum (V_T+w) of terminal fall speed of the hydrometeors and vertical air velocity. In order to estimate the terminal fall velocity, a statistical approach has been recently proposed in the case of frontal cyclones and weakly-precipitating ice clouds (Protat et al., 2003). It consists in developing statistical relationships between terminal fall velocity and radar reflectivity, which can be written as $V_T=aZ^b$, where Z is expressed in $\mathrm{mm}^6\mathrm{m}^{-3}$, and V_T in ms^{-1}. Within weakly-precipitating clouds, the vertical air motions are generally small, even at small scales of motion, as opposed to the case of convective systems. In any case, however, the vertical air motions are not negligible with respect to the terminal fall speed. For a long time span however (a few hours), the mean vertical air motions should vanish with respect to the mean terminal fall speed, which is much less fluctuating. A statistical power-law relationship between the terminal fall speed and radar reflectivity may therefore be derived from this statistical approach. This hypothesis has been recently validated in the case of frontal cyclones sampled during FASTEX (Protat et al., 2003). A more thorough validation of this assumption will be performed in a near future using high-resolution numerical simulations of cirrus clouds. Once the V_T-Z statistical relationship is obtained the radar reflectivity is easily translated into terminal fall velocity at each radar gate. Another approach, proposed by Matrosov et al. (2002) consists in simply averaging the Doppler measurements over 20–30 min and assume that the mean vertical air velocity is negligible with respect to the mean terminal fall velocity. The shortcoming of this approach is that the hypothesis is assumed on a much smaller time span, but a significant advantage is that it is less sensitive to a change in the cloud microphysics. We will soon investigate the conditions under which a method is better than the other, but for the present paper we have retained the first approach.

2.3 Principle of the radar retrieval method

Once the ice density is estimated from the σ_D/V_T ratio, then all the relationships of Sect. 2.1 are computed using this ice density. The remaining unknowns to access the ice cloud properties are N_0^* and D_m. In order to estimate D_m, we have developed relationships between V_T and D_m, parameterized by the retrieved ice density. This is a major difference with the Matrosov et al. (2002) method, who assumed that density is that of solid ice to compute a single V_T-D_m relationship for all types of ice clouds.

Regarding N_0^*, there is an analytical relationship between N_0^*, D_m, and the radar reflectivity Z when Rayleigh scattering and an analytical shape of the normalized PSD are assumed. In Mie scattering, which is very likely to occur for radars at 95 GHz, there is no simple analytical relationship between these quantities, but this relationship can be simply derived as a look-up table. As a result, N_0^* can be estimated from Z and the D_m retrieved from V_T. The uncertainties arising from this method have been assessed by Delanoë et al. (2004), using an extensive database of airborne in-situ microphysical measurements already mentioned in Sect. 2.1. Fig. 2 shows the N_0^* retrieved using the analytical shape of Eq. (3) as a function of the true N_0^* computed from the true PSDs for all the database. The obtained mean error and standard deviation on N_0^*, which includes both the errors due to the assumption on shape and those due to the $N_0^*-D_m-Z$ approximation, are of about −2.5% and 15.7%, which by construction translates into roughly the same errors on the retrieved cloud parameters (Delanoë et al. 2004). This error is

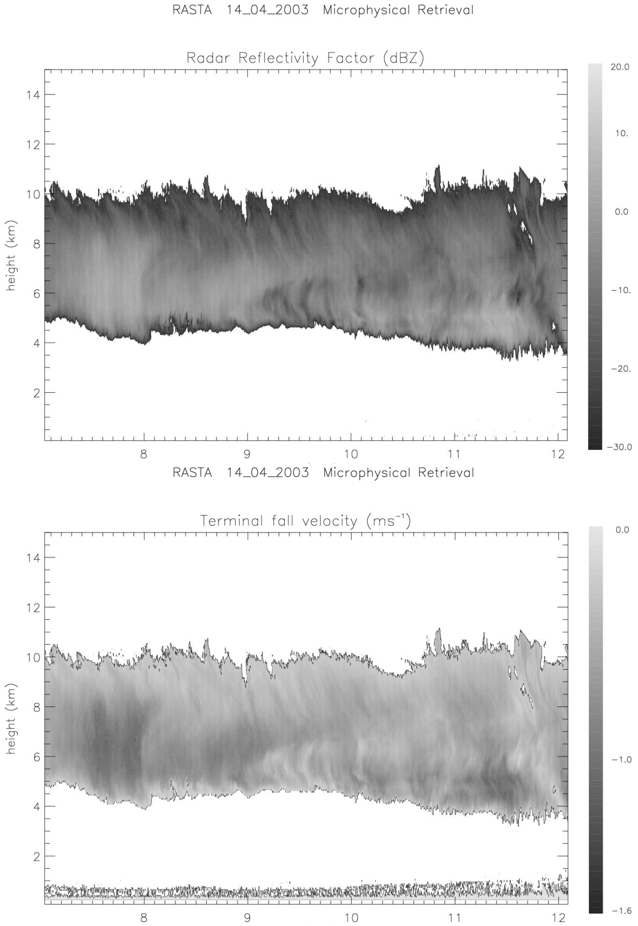

Fig. 3. Time-height cross-sections of Z and V_T. These data have been collected by the RASTA 95 GHz cloud radar over the SIRTA, Palaiseau, France, on 14 April 2003.

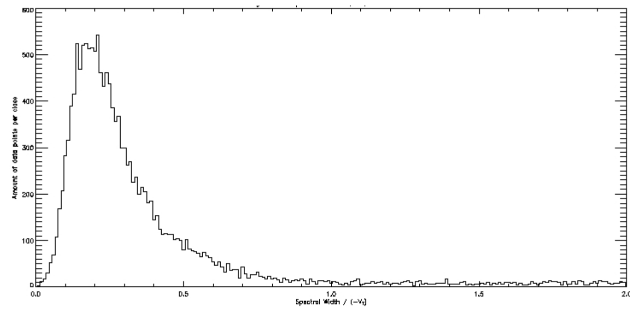

Fig. 4. Statistical distribution of the σ_D/V_T ratio for the 14 April 2003 ice cloud.

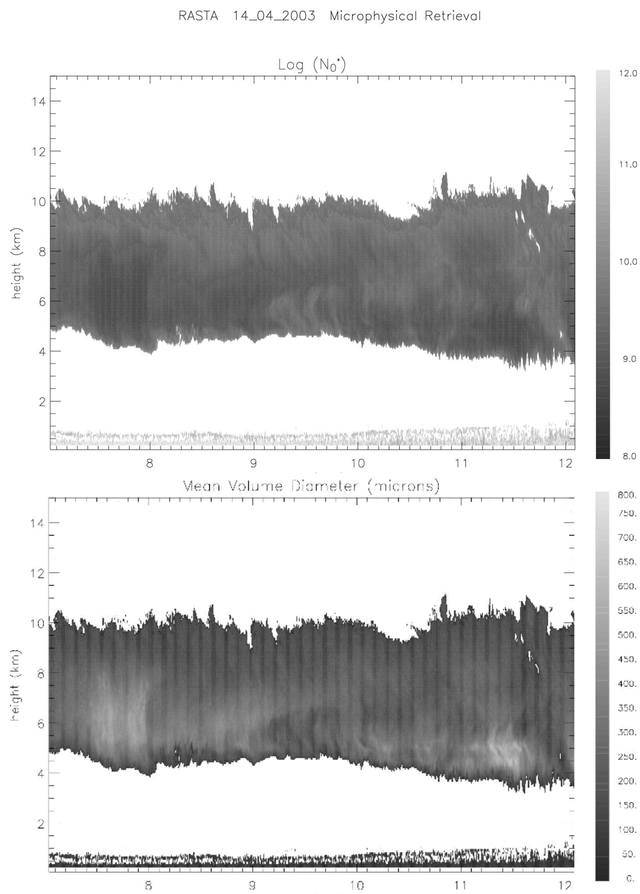

Fig. 5. Same as Fig. 3, but for the retrieved N_0^* and D_m.

mostly due to the assumption on the normalized PSD shape. A global error analysis of the method has not been conducted yet, but it is foreseen that errors on IWC, effective radius, and visible extinction should not exceed 20–25%, if the ice density is correctly determined. Using the retrieved N_0^* and D_m, the cloud properties can be finally retrieved from Eq. (2) (IWC), Eq. (5) (R_e), and the relationship between Z/N_0^* and α/N_0^* discussed in Sect. 2.1. In the next section, the method is applied to the case of a thick prefrontal ice cloud.

3 Illustration of ice cloud retrieval using the radar-only method

The method described in Sect. 2 has been applied to continuous Doppler cloud radar measurements at 95 GHz collected in the frame of the European CloudNET project over the SIRTA (Site instrumented for cloud studies in Palaiseau, France). The case shown here as an illustration is that of a thick midlatitude prefrontal ice cloud. A backscatter lidar was also operating at that time, but the optical depth of the ice cloud was such that only few hundred meters of the cloud were penetrated by the lidar until complete extinction. This case is therefore a good illustration of the complementarity between the radar method and the radar-lidar method, since the radar method allows in this particular case to explore the upper part of the ice cloud that cannot be reached by the lidar. The radar reflectivity and the terminal fall velocity retrieved by the method described in Sect. 2.2 are given in Fig. 3. The radar reflectivity is such that we expect Mie scattering to occur in significant parts of the ice cloud, which is accounted for in the radar method. The first step is to derive information on the ice density from the radar spectral width to terminal fall velocity ratio. During this study we have obtained that

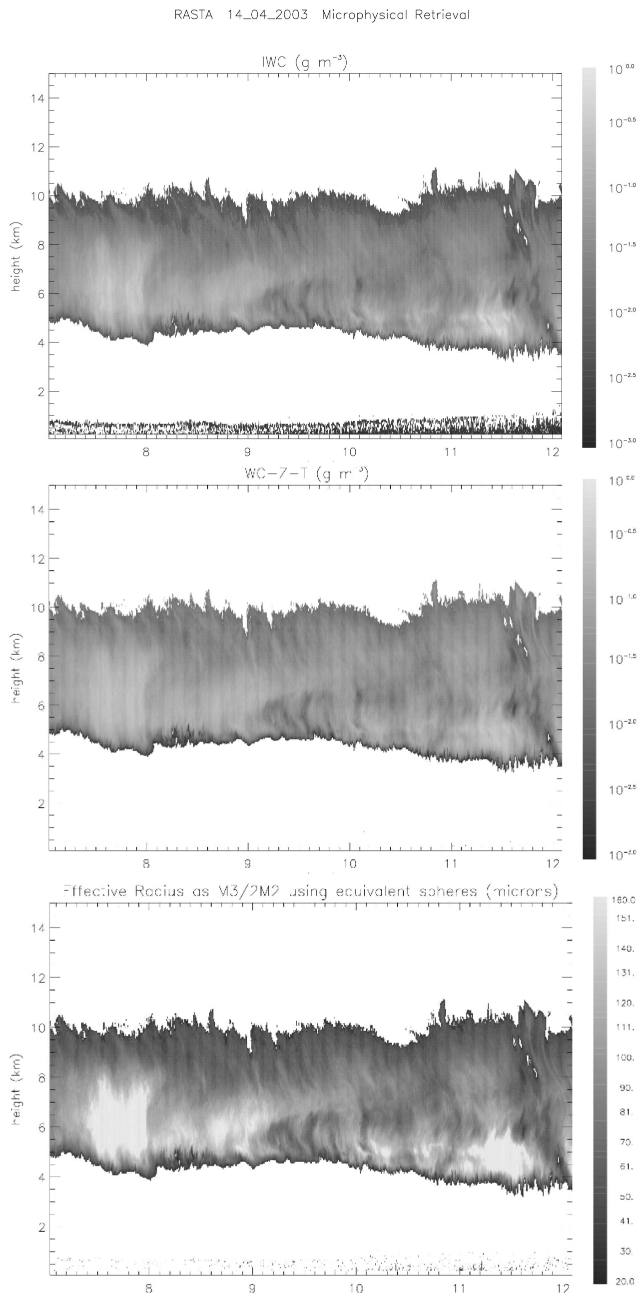

Fig. 6. Same as Fig. 3, but for the retrieved IWC using the present radar method, IWC using the IWC-Z-T method, and the retrieved R_e.

our real-time calculation of spectral width was not accurate for low signal-to-noise ratios, but correct inside the cloud. This problem has been recently fixed on our radar. Owing to this calculation problem, we have chosen as a first step to derive a single information on ice density rather than a vertically-resolved one, which represents already a large improvement over other methods.

Figure 4 shows the statistical distribution of the σ_D/V_T ratios (the areas of small signal-to-noise ratios are included, which tends to make this distribution wider). A peak is clearly obtained for values ranging from 0.15 to 0.2, which

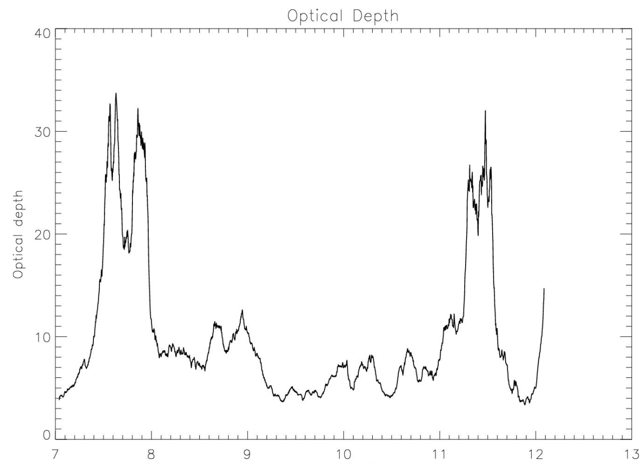

Fig. 6d. Same as Fig. 3, but for the retrieved cloud optical depth.

corresponds to B coefficients ranging from 0.4 to 0.5. A review of the literature shows that such an exponent corresponds fairly well to the exponent of the relationship proposed by Locatelli and Hobbs (1974) for aggregates, while it does not correspond at all to typical values for other typical crystals such as hexagonal columns, bullet rosettes or planar polycrystals, which are habits commonly assumed in radar-lidar retrievals (see summary of B coefficients in Matrosov et al. 2002). The corresponding Locatelli and Hobbs (1974) density-diameter relationship has therefore been used in what follows.

Figure 5 shows the N_0^* and D_m retrieved using the radar method described in Sect. 2. It is to be noted that these values and variations as a function of the radar reflectivity and terminal fall velocity seem to be fairly consistent with those derived by Delanoë et al. (2004) from the database of airborne in-situ microphysical measurements. As an example, the smallest N_0^* are associated with the largest reflectivities, which reflects the decrease of particle concentration and increase of particle size inside the cloud with respect to the cloud edges, as a result of aggregation going on. Finally, once N_0^* and D_m are retrieved, we can access the ice cloud properties.

Figure 6 shows the ice water content and effective radius retrieved for the same cloud as previously. The could optical depth retrieved from the visible extinction is also given in Fig. 6d. In order to compare the retrieved ice water content to another radar method, we have also computed ice water content using the so-called $IWC-Z-T$ method proposed by Liu and Illingworth (2000), which expresses in a simple manner IWC as a function of radar reflectivity and temperature. Temperature was provided in our case by a radiosounding launched at 11:20 UTC 15 km away from the radar. The expected uncertainty of such a method is of about 70–100%.

The comparison of the IWCs is in the overall encouraging. A closer inspection reveals though that the radar method produces much larger ice water contents in the areas of strong reflectivities than the $IWC-Z-T$ method, which can be explained by the fact that the data sample used to build the $IWC-Z-T$ method did not include such large reflectivities. It is also seen that the radar method produces generally smaller IWCs at cloud edges. Effective radii are in the correct range for such thick ice clouds (from 20 to 160 μm), but there is no independent estimate of effective radius available for validation. In a near future, this radar retrieval method will be compared with the radar-lidar retrieval method in the cloud areas sampled by both the radar and the lidar. This will be conducted using the whole CloudNET radar-lidar database collected from three European instrumented sites (Chilbolton, UK, Cabauw, Netherlands, Palaiseau, France).

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