RETRIEVALS OF CLOUD CONTENT AND PARTICLE CHARACTERISTIC SIZE USING NOAA ENVIRONMENTAL TECHNOLOGY LABORATORY CLOUD RADARS

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1. INTRODUCTION

The National Oceanic and Atmospheric Administration's (NOAA) Environmental Technology Laboratory (ETL) has been actively involved in radar cloud research for more than a decade. One of the first millimeter wavelength radars dedicated almost entirely to studies of non-precipitating and weakly precipitating clouds was the NOAA/K radar (Pasqualucci et al., 1983, Kropfli et al., 1990). This is a transportable, polarization-agile, 8.6 mm (K_a-band) radar with full scanning and Doppler capabilities. The NOAA/K radar was part of many recent cloud field experiments, and, in a way, many exciting results obtained with this radar inspired cloud radar program developments in other institutions.

One example of such developments is the cloud radar part of the Atmospheric Radiation Measurement (ARM) Program of the United States Department of Energy (DOE). ETL manufactured several vertically-pointed 8.6 millimeter wave cloud radars (MMCR) for the operational use at three ARM cloud and radiation testbed (CART) sites. These radars are designed for unattended use, they have full Doppler capability and their possible polarization upgrade is considered (Moran et al., 1998). One MMCR has been available at ETL for the use in different field experiments.

A number of cloud retrieval algorithms that utilize the radar data have been developed in ETL. The focal point of these algorithms is radar data; however, most of them also use collocated measurements from other remote sensors such as IR and microwave radiometers and/or lidars. Though ETL retrieval algorithms were designed primarily for the ground-based measurements, some of these algorithms, with certain adjustments, can also be used with measurements taken from satellite instruments. In this paper, we give a brief review of some ETL radar group cloud retrieval algorithms, emphasizing those algorithms which can be used for spaceborne cloud profiling radar measurements.

2. RETRIEVING LAYER-MEAN PARAMETERS

One of the first retrieval algorithms developed in ETL for ice clouds (Matrosov et al. 1992) allows estimation of mean-layer particle characteristic size (such as mean, effective, or median) and the value of ice water path (IWP, i.e., the vertically integrated cloud ice water content-IWC) from the layer mean radar reflectivity $\overline{Z_e}$ and an estimate of cloud optical thickness τ. Cloud optical thickness in the atmospheric window region is usually obtained from measurements of a vertically-pointed IR radiometer (λ~ 10-11.5 μm). The main idea of this approach is based on $\overline{Z_e}$ and τ being approximately proportional to the sixth and the second vertically integrated moments of the cloud particle size distribution (PSD). The mean-layer algorithm was subsequently refined and generalized (e.g., Matrosov et al., 1998) accounting for the size-dependent particle density, multiple scattering effects, and cloud temperature gradients. An example of the application of this algorithm is shown in Fig. 1 for the ice cloud case observed during the Surface Heat Budget of the Arctic Experiment (SHEBA). A similar layer-mean algorithm is used by ARM (Mace et al., 1998).

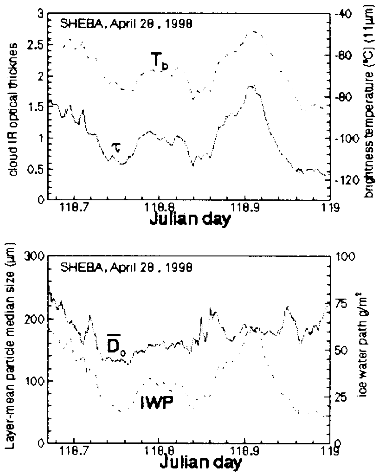

Fig 1 Results of the retrieval of layer-mean ice cloud parameters

Estimates of liquid water path (LWP) for warm clouds can be obtained from two-channel microwave radiometer measurements and droplet characteristic size in such clouds is strongly correlated with radar reflectivity in the absence of drizzle (Frisch et al., 2000).

3. RETRIEVALS OF VERTICAL PROFILES OF CLOUD PARAMETERS USING DOPPLER DATA

The knowledge of the vertical profiles of cloud microphysical parameters is necessary for adequate representations of clouds in climate models. Several retrieval algorithms have been developed at ETL for profile retrievals. For ice clouds, measurements of vertical Doppler velocities, V_D, could be used for estimating vertical profiles of the reflectivity-weighted particle fall velocity, V_z (Orr and Kropfli 1999). The reflectivity (Z_e) and fall velocity (V_z) profiles with an additional estimate of cloud optical thickness (τ) from IR measurements can then be used to retrieve profiles of particle total number concentration (N) and particle median size (D_o) independently. This is achieved by an iterative solution of a non-linear system of algebraic equations (Matrosov, 1997):

$$\tau = \sum_i k_1(\rho,n,D_{oi})N_iD_{oi}^2\Delta h_i$$
$$Z_{ei} = k_2(\rho,n,D_{oi})N_iD_{oi}^6$$
$$V_{zi} = A\, k_3\,(\rho,n,D_{oi})D_{oi}^B \qquad (1)$$

Here coefficients k_1, k_2, and k_3 depend on assumptions about particle density ρ, and the order of the gamma function size distribution n. The summation is over all range gates within a cloud, and the subscript i refers to the values of the median size (D_o) and the total number concentration (N) in an individual radar range gate. The coefficient A and the exponent B relate particle size to its fall velocity, and k_3 describes the transition between an individual particle to the particle ensemble.

The natural variability of the coefficient A is more than one order of the magnitude, so it is regarded as unknown. However, there is a relation between A and B (Matrosov and Heymsfield 2000) which reduces the number of unknowns for each radar beam to $2L+1$, where L is a number of radar range gates within the cloud (i.e., L values of N, L values of D_o and a value of A). The number of input parameters for the system (1) is also $2L+1$ (i.e., L values of Z_e, L values of V_z and a value of τ) which allows independent retrievals of all unknowns.

The iterative solutions usually converge after 2-3 iterations. As soon as the vertical profiles of N and D_o are retrieved, one can also calculate the vertical profile of IWC.

Droplets in non-precipitating liquid water clouds are very small and their fall velocities are usually negligible compared with vertical air motions. This fact prevents estimating droplet fall velocities in the way it can be done for ice cloud particles which usually are much larger than droplets. However, the fact that microwave radiometer measurements provide a direct estimate of LWP makes liquid water cloud retrievals relatively easier than in a case of ice clouds (whenever that contributions from drizzle size drops and boundary layer "insects" are absent or negligible).

ETL liquid water cloud retrievals developed by Frisch et al. (1995, 1998) assume that the width of the PSD and the total droplet number concentration, N, do not change with height. The values of LWC in each radar range gate (i) can be calculated then from:

$$\mathrm{LWC}_i = \mathrm{LWP}\, Z_{ei}^{0.5} / \left(\sum_i Z_{ei}^{0.5}\Delta h_i\right). \qquad (2)$$

This retrieval does not depend on the absolute radar calibration or on values of the PSD width and N if there is an approximate proportionality between the sixth moment of the PSD and the square of the third moment of the PSD. After the vertical profile of LWC is retrieved, the profile of the drop characteristic size (e.g., effective radius) can be calculated for an assumed value of the PSD width. Droplet size retrievals, though, depend on the absolute radar calibration.

Both ETL ice and liquid water cloud retrieval algorithms have been extensively used in different cloud field programs held in recent years. On a number of occasions, these algorithms were verified by comparisons of the retrieved values of cloud parameters with their estimates from in situ aircraft sampling. During these comparisons, every effort was made for the best possible collocation of the remote and direct sampling.

These comparisons (Matrosov et al., 1998, Frisch et al., 1999) were rather encouraging. For ice clouds, a typical standard deviation between remotely measured and directly estimated values of D_o were about 30% for IWC the corresponding deviations were about 50%. Such an agreement should be considered good given the uncertainties of both types of measurements and the natural variabilities of D_o and IWC (more than one and four orders of magnitude, respectively). For the liquid water clouds, the relative standard deviation between remote and direct measurements is about 20% (Frisch et al., 1999). Such an agreement was obtained for the range of LWC from about 0.05 to 0.6 g m^{-3}. A part of the reason that the agreement for water content was better than for ice content is that in the case of liquid clouds the normalizing value of LWP is available directly.

4. TUNED REGRESSION RETRIEVALS FOR ICE CLOUDS

One of the most difficult parts of the ice cloud profiling approach described earlier is the use of Doppler velocity measurements. Since the Doppler data need to be averaged in a certain manner (Orr and Kropfli, 1999) to estimate particle fall velocities, this approach is applicable when vertical air motions are relatively weak, and no strong turbulence is present in clouds. These conditions are not satisfied all the time; besides, sometimes the Doppler information is contaminated or not available at all.

One of the practical examples when Doppler measurements could not be effectively used for retrievals is the year long (October 1997-October 1998) SHEBA experiment. During this experiment, the ETL MMCR radar was deployed on board of an icebreaker which was drifted in the Arctic basin to the north of Alaska. Due to ice movements, the direction of the radar beam at any particular time was not perfectly aligned with respect to the vertical. A typical deviation of the beam from the zenith direction was about 1^0-1.5^0. Because of this misalignment, Doppler velocity measurements were contaminated by the horizontal winds which are much stronger than typical ice cloud particle fall velocities. This contamination was difficult to account for since horizontal winds were often changing significantly between two consecutive rawinsonde launches. Nearly vertical IR and microwave measurements, however, were available to complement radar data for the most part of the experiment.

A relatively simple ice cloud profiling algorithm was developed at ETL for the SHEBA instrument configuration (Matrosov, 1999). This algorithm is based on "tuning" the coefficients of power-law empirical IWC-Z_e relations for each beam of the radar data. This procedure allows retrievals of IWC profiles with consecutive calculations of particle characteristic size profiles, D_o.

It can be shown (e.g., Atlas et al. 1995) that radar reflectivity of clouds can be expressed in terms of IWC and D_o in the following way:

$$Z_e = G \, \mathrm{IWC} \, D_o^3. \tag{3}$$

The parameter G depends on details of the PSD, and, for ice clouds, also on the particle density and shape. For quasi-spherical particles, the exponential PSD and a typical particle bulk density dependence suggested by Brown and Francis (1995)[ρ (g cm^{-3}) $\approx 0.07\, D^{-1.1}$ for D>0.1 mm], G can be approximated as:

$$G \approx 7.5 \cdot 10^{-5} D_o^{-1.1} \tag{4}$$

if Z_e is in mm^6/m^3, IWC is in g m^{-3}, and D_o is in μm. The dependence of G on D_o in (4) is a proxy for its dependence on ρ. This dependence is much stronger than the ones on PSD details and on particle shape.

It can be seen from (3) and (4) that Z_e is mostly determined by two parameters: IWC and D_o. Note that the variabilities of Z_e due to natural changes of both IWC and D_o are of approximately the same range (3-4 orders of magnitude). If IWC and D_o are independent, the retrieval of either IWC or D_o from one measurements of Z_e is impossible. However, as in situ and independent remote measurements of IWC or D_o (using reflectivity and Doppler measurements together) show there is a noticeable correlation between these cloud parameters. This correlation allows constructing empirical relations of a type:

$$\mathrm{IWC} = a_1 Z_e^b \tag{5a}$$
$$D_o = a_2 Z_e^c \tag{5b}$$

There has been a number of experimental studies suggesting different values for a_1 and b. Most of them were based on in situ data when IWC and Z_e were calculated from particle samples. These studies suggest a wide range of values for a_1 while the suggested values of b are usually in the range from about 0.55 to 0.75. The "tuned" regression algorithm for the SHEBA data set requires adjusting values of the coefficient a_1 for each radar beam of reflectivity measurements. The corresponding value of a_1 is found from:

$$a_1 = \mathrm{IWP} \, / \, \sum_i Z_{ei}^b \, \Delta h_i \, . \tag{6}$$

The value of b used in (6) could be assumed to be about 0.65 which is in the middle of its variability interval. It was shown (Matrosov 1999), however, that better results can be obtained if b is assumed to vary linearly diminishing from about 0.7 near the cloud base to about 0.6 at the cloud top.

The normalizing value of IWP in (6) is estimated using the layer mean approach from the layer mean radar reflectivity $\bar{Z}_e$ and an IR estimate of cloud optical thickness τ:

$$\mathrm{IWP} = k_4 \, \bar{Z}_e^{0.25} \, \tau^{0.75}, \tag{7}$$

where k_4 depends on details of the PSD. In a way, this algorithm is similar to the one for retrieving profiles of LWC with an exception that the vertical integral normalizing value of IWP is not available readily (like LWP is available from microwave radiometer measurements) but it is estimated from a combination of different measurements with some uncertainty.

After the vertical profile of IWC is retrieved from the a "tuned" regression specifically tailored for a given vertical profile of Z_e measurements, a corresponding vertical profile of D_o can be calculated using (3) and (4).

The "tuned" regression retrieval algorithm was applied for all ice cloud cases observed during the year-long SHEBA observations. A case of April 29, 1998 was of a particular interest since coincident aircraft in situ measurements were available. Fig. 2 shows a time-height cross-section of measured radar reflectivities for this case.

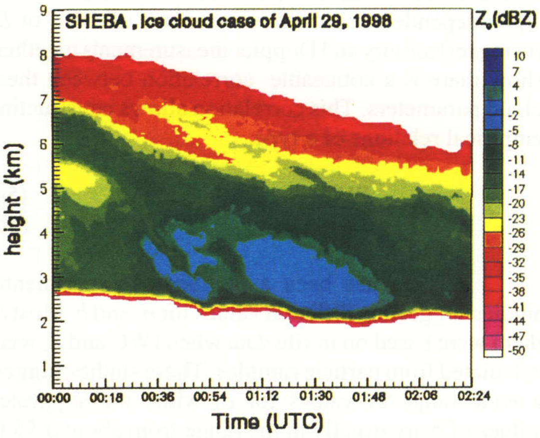

Fig. 2. Measured radar reflectivities

In spite of being very thick geometrically, this cloud remained optically semi-transparent, which allowed estimations of cloud optical thickness from IR data. Figures 2 and 3 show results of the retrieval of the IWC and the particle characteristic size (expressed in terms of the mean size D_{mean}) using the algorithm described above. Note that for typical ice cloud PSD mean size is usually significantly smaller than the median size (i.e., D_o).

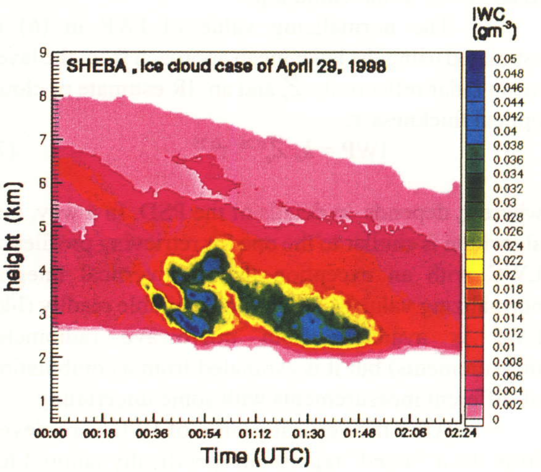

Fig. 3. Retrieved cloud ice water content values

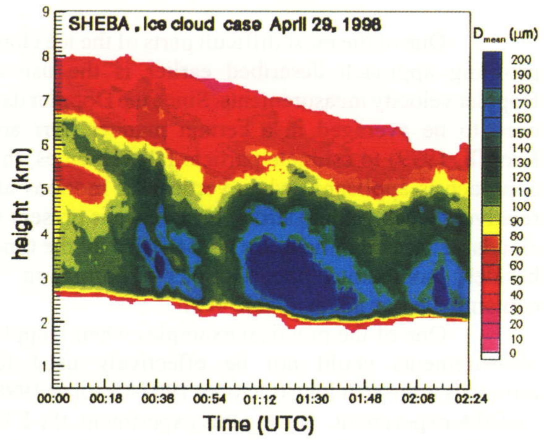

Fig. 4. Retrieved particle mean size values

The Canadian CV-580 aircraft was performing a spiral decent in this cloud above the SHEBA ground site from 00:00 to about 00:15 UTC. The aircraft in situ measurements (data courtesy of A. Korolev and G. Isaac of the Canadian Atmospheric Environmental Service) were used to calculate IWC and particle mean sizes. Comparisons of remote and in situ derived cloud parameters are shown in Fig.5. The retrieval profiles are shown for the whole time of the aircraft descent in 4 minute intervals.

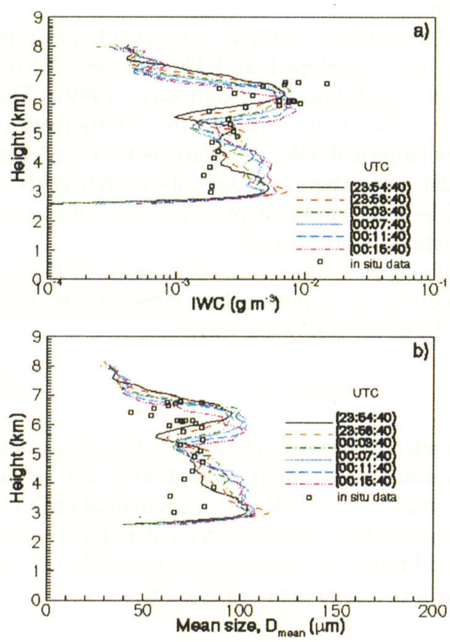

Fig.5. Comparisons of in situ and remote measurements of IWC (a) and mean particle size (b).

5. RETRIEVALS OF CLOUD PARAMETERS BASED ON REFLECTIVITY ONLY

The described earlier ETL algorithms use a multi-sensor approach. However, on some practical occasions, like a minimum CLOUDSAT mission, only radar reflectivity measurements might be available. This dictates a need for the algorithms which use reflectivity only data. Such algorithms are inevitably based on one-parameter relations of the type given by (5), however, in this case, "tuning" parameters of these relations, based on coincident measurements of other remote sensors, would not be possible.

5.1. Effective radius - reflectivity relations for liquid water clouds

If drizzle-size drops are absent, the warm cloud PSD can be satisfactorily described by the three parameter log-normal distribution (Frisch et al., 1995). Integrating this distribution, one can get the following expressions for the reflectivity, Z_e and LWC:

$$Z_e = 2^6 N_o r_e^6 \exp(3\sigma^2), \quad (8)$$

$$\mathrm{LWC} = 1.33\, \pi \rho N_o r_e^3 \exp(-3\sigma^2), \quad (9)$$

where N_o, r_e, ρ, and σ are the total droplet concentration, the effective radius, the water density, and the PSD width, respectively. The units of σ are such that in PSD droplet sizes are expressed in microns. If, analogous to (3), the basic equation for the radar reflectivity is expressed in terms of LWC and r_e, one can show that for warm clouds the coefficient G depends only on the PSD width. If Z_e is in mm^6/m^3, LWC is in g m^{-3}, and r_e is in µm:

$$G \approx 1.53 \cdot 10^{-6} \exp(6\sigma^2). \quad (10)$$

From (8) the expression for r_e follows as:

$$r_e = [2 \cdot \exp(0.5\sigma^2) N_o^{1/6}]^{-1} \cdot Z_e^{1/6} \quad (11)$$

From in situ aircraft measurements in Oklahoma stratus clouds Frisch et al. (2000) found that the mean value of σ is about 0.32 with a standard deviation of about 0.07, and the mean value of $N_o^{1/6}$ is about 2.5 (i.e., $N_o \approx 244$ cm^{-3}). It was also found that the variability of the total droplet concentration significantly reduces if $Z_e > 0.001$ mm^6/m^3 (i.e., -30 dBZ which is approximately equal to the sensitivity limit of the proposed CLOUDSAT radar). For these mean Oklahoma conditions, (11) reduces to:

$$r_e(\mu\mathrm{m}) = a_2 Z_e^c \approx 19 \cdot Z_e^{1/6} (\mathrm{mm}^6/\mathrm{m}^3). \quad (12)$$

Here we introduced parameters a_2 and c for the cloud particle characteristic size - reflectivity relations as in (5b) but for the case of liquid water clouds when characteristic size is expressed in terms of the effective radius.

The relation (12) is very close to the one suggested by Fox and Illingworth (1997) for marine stratus clouds:

$$r_e(\mu\mathrm{m}) \approx 23 \cdot Z_e^{0.176} (\mathrm{mm}^6/\mathrm{m}^3). \quad (13)$$

Note that the empirical exponent of 0.176 in (13) is very close to the theoretical value of 1/6 in (12). In spite of the fact that (12) and (13) were obtained using different approaches and for different geographical conditions, the coefficients of these relations (i.e., 19 and 23) are quite close.

This relatively small variability of r_e - Z_e relations can be understood by analyzing the theoretical expression (11). It can be seen from (11) that significant variations in N_o and σ result in relatively small variations of the coefficient a_2 thanks to the fact that this coefficient is sensitive to the 6-th root of N_o and the term $\exp(0.5\sigma^2)$ does not change much for the dynamic range of the variability in the PSD width σ. The variability in σ in the range from 0.25 to 0.39 which is characteristic to the Oklahoma clouds (Frisch et al. 2000) results in only 5 % changes in the coefficient a_2 in (12). Significant changes in the total droplet concentration from 100 cm^{-3} to 500 cm^{-3} result in only 25% variability in a_2.

This sensitivity analysis indicate that, with the absence of drizzle, the r_e - Z_e relations for warm clouds are quite stable. The presence of drizzle would contaminate estimates of cloud droplet effective radius from reflectivity only measurements. Some ways to identify this presence are to use the reflectivity threshold (Frisch et al. 1995) or the shape of the reflectivity profile in cloud (Fox and Illingworth 1997).

5.2. LWC- reflectivity relations for liquid water clouds

For the log-normal PSD, the radar reflectivity can be expressed either in terms of LWC, r_e and σ (as given by (3) and (10) if r_e is substituted for D_o) or in terms of N_o, LWC and σ. The latter way of this expressing is more convenient for deriving LWC-Z_e relations:

$$Z_e = (3.66/N_o) \exp(9\sigma^2)\, \mathrm{LWC}^2. \quad (14)$$

or

$$\mathrm{LWC} = (N_o^{0.5}/1.91) \exp(-4.5\sigma^2)\, Z_e^{0.5} \quad (14a)$$

For average Oklahoma warm cloud droplet distribution width (σ =0.32) (14a) can be re-written as:

$$\mathrm{LWC}(\mathrm{g\ m^{-3}}) = a_1 Z_e^b \approx (N_o^{0.5}/3)\, Z_e^{0.5}. \qquad (15)$$

Here N_o is in cm^{-3} and Z_e are in mm^6/m^3. In (15) we used the same coefficient notations for the LWC-Z_e power-law relation as in equation (5a). A LWC - Z_e relation suggested by Liao and Sassen (1994) for stratus clouds:Z_e= $(3.6/N_o)\mathrm{LWC}^2$ can be written a similar to (15) way as:

$$\mathrm{LWC}(\mathrm{g\ m^{-3}}) = (N_o^{0.56}/2)\, Z_e^{0.56}. \qquad (16)$$

As can be seen from Fig.6 for an average droplet concentration in continental stratus clouds ($N_o \approx 250$ cm^{-3}) LWC values from (15) are smaller than those from (16) by 30%-40% for typical reflectivities of non-precipitating stratus clouds.

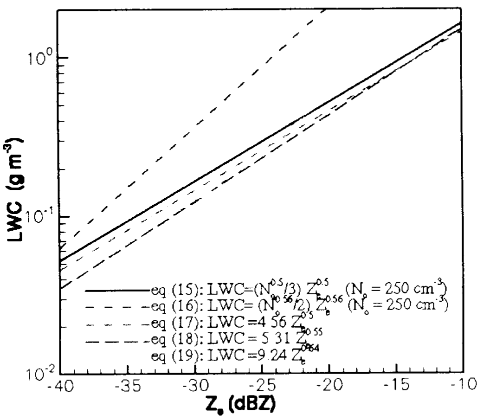

Fig.6. Comparisons of different LWC-Z_e regressions

Since both LWC and Z_e are proportional to the droplet concentration and this concentration tends to be constant with height in relatively thin stratus clouds (Frisch et al. 1995), the deviation of the exponent b in LWC-Z_e relations from the its theoretical value of 0.5 causes an apparent unit conflict in (16). It should be mentioned , however, that (16) was found empirically using numerical simulations with an adiabatic cloud model. The deviation of b from 0.5 also can be a result of some correlation between cloud particle concentration and radar reflectivity (which is, probably, the case with ice clouds where $b \approx 0.65$).

As it can be seen from Fig. 6, (15) agrees favorably with earlier relations suggested by Atlas (1954):

$$\mathrm{LWC}\ (\mathrm{g\ m^{-3}}) = 4.56\, Z_e^{0.5} \qquad (17)$$

and Sauvageot and Omar (1987):

$$\mathrm{LWC}\ (\mathrm{g\ m^{-3}}) = 5.31\, Z_e^{0.55}. \qquad (18)$$

For the range, -30 dBZ< Z_e< -20 dBZ, there is also a decent agreement between (15) and the relation suggested by Fox and Illingworth (1997):

$$\mathrm{LWC}\ (\mathrm{g\ m^{-3}}) = 9.24\, Z_e^{0.64}. \qquad (19)$$

Note, however, that (19) was derived empirically for marine stratus clouds, while the good agreement is reached when a total cloud droplet concentration value (250 cm^{-3}) typical for continental stratus clouds was used for the relation (15) in Fig. 6.

Comparing (14b) and (11) indicates that the changes in the coefficient a_1 of LWC-Z_e relations (15) due to the variability in PSD details (i.e., σ) and the droplet total concentration, N_o is much greater than corresponding changes in the coefficient a_2 of r_e-Z_e relations. Simple sensitivity calculations show that the variability in a_1 when σ changes from 0.25 to 0.39 and N_o changes from 100 cm^{-3} to 500 cm^{-3} is about 40% and 70%, respectively. The corresponding changes in a_2 were 5% and 25% only.

This sensitivity displays the need of appropriate a priori assumptions about σ and N_o for the use in the coefficients of LWC-Z_e relations. If such assumptions are made, retrievals of LWC using radar reflectivity only measurements can yield good results. This is illustrated in Fig. 7 where, for one of the Oklahoma stratus cloud observations, the time series of LWP values obtained by integrating results from 3 different LWC-Z_e relations and from microwave radiometer data are displayed.

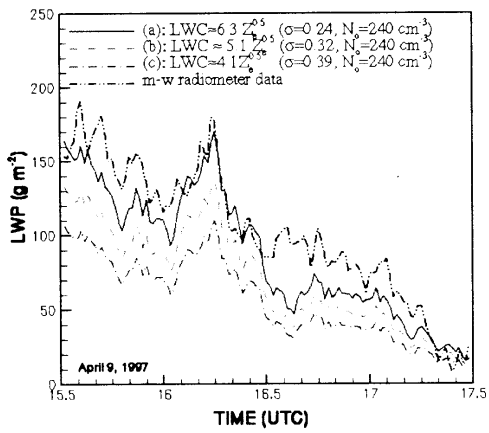

Fig.7. Comparisons of stratus cloud LWP estimates from radar (a), (b), (c) and from microwave radiometer

From aircraft in situ measurements during this case it was known that the mean values of σ and N_o in cloud parts which were sensed by the radar were approximately 0.24 and 240 cm^{-3}. The radar derived LWP curve (a) corresponding to these values demonstrates a good agreement with microwave radiometer measurements of LWP.

5.3. Cloud microphysical parameters - reflectivity relations for ice clouds

For ice clouds, there is more uncertainty in theoretical calculations of radar reflectivity (compared to liquid water clouds) because of the variability in ice particle shapes and bulk density. This results in more uncertainty in the coefficients of the power-law relations (5).

As was mentioned above, ETL participated in many recent cloud field experiments (e.g., FIRE-II, ASTEX, 1995 Arizona Program, several ARM cloud experiments in Oklahoma) during the past decade. During these experiments, a number of high-quality ice cloud cases were observed with the Doppler NOAA/K radar and the suite of IR and microwave radiometers. Applying the Doppler radar - radiometer profiling algorithm described in section 3 to these ice clouds allowed independent retrievals of vertical profiles of IWC and D_o. These microphysical retrieval data and coincident measurements of radar reflectivity were used then to derive IWC - Z_e and D_o - Z_e relations on a case-by-case basis. The dynamic range of the IWC - Z_e relations found using this approach were in general agreement with variability in empirical IWC - Z_e relations derived from in situ data. Some results of comparisons and discussions of the variability in these relations were discussed by Matrosov (1997).

The average IWC-Z_e and D_o -Z_e relations for this set of ice cloud observations are:

$$\text{IWC (g m}^{-3}) \approx 0.125 \cdot Z_e^{0.62}, \qquad (20)$$

$$D_o\,(\mu\text{m}) \approx 420 \cdot Z_e^{0.18} \,. \qquad (21)$$

The exponent 0.62 in (20) is in good agreement with results of many empirical derivations of IWC-Z_e relations based on in situ data. These derivations usually yield values of b around 0.65. The exponent 0.18 in (21) is rather close to 1/6 as in the theoretical droplet effective size-reflectivity relation (11). Case-to-case variations in IWC-Z_e and D_o - Z_e relations indicate a greater relative variability in the coefficients than in the exponents. Relations between a characteristic size of ice cloud particles and radar reflectivity exhibit less variability than those between cloud content and reflectivity. As for liquid water clouds, this result is mostly due to a smaller sensitivity of the parameters in the characteristic size-reflectivity relations to details of the PSD and to the particle total concentration.

Equations (20) and (21) represent the average relations for ice clouds observed in the field experiment mentioned above. Averaging was done for all ice clouds regardless of observation conditions. For an effective use of such relations on a global basis, their parameters, probably, need to be adjusted depending on conditions of a particular observation. The most obvious criteria for such adjustments would be geographical locations, seasons and cloud heights. Since the exponents (c and b in (5)) are relatively stable, these adjustments should primarily concern the coefficients (i.e., a_1 and a_2 in (5)).

An example of such possible adjustments, can be illustrated using the SHEBA results. The tuned regression algorithm described earlier allows changing the coefficient a_1 for each radar beam using cloud optical thickness estimate from IR measurements and the layer-mean radar reflectivity. The mean value of this coefficient for all ice cloud experimental cases in SHEBA was about 0.08 which is smaller than the mean value of 0.125 in (20) obtained from cloud cases observed with NOAA/K. In part, this difference can be explained by the fact that often the low level ice clouds in the Arctic consist of larger particles at smaller concentrations compared to high tropospheric cirrus clouds for which the average relations (20) and (21) were obtained. These specific conditions cause an increase in a_2 and a decrease in a_1.

Since the pair of relations (20) and (21) relates one measurement of radar reflectivity to two different cloud microphysical parameters (which is possible only due to some inherent correlation between IWC and D_o), these relations should not contradict each other in a sense that estimated values of IWC and D_o should still satisfy the basic equation (3). Such a mutual consistency is present for (20) and (21). This can be illustrated by substituting the value of the median particle size, D_o from (21) and the value of the parameter G from (4) in the basic equation for the radar reflectivity (3). These substitutions yield the following expression for IWC:

$$\text{IWC (g m}^{-3}) \approx 0.13 \cdot Z_e^{0.66}, \qquad (22)$$

which is very close to (20). Given some inevitable uncertainty in the parameter G due to particle shape, bulk density and details of PSD (Matrosov 1999), relations (20) and (22) could be considered practically identical.

6. CONCLUSIONS

Over the past decade the NOAA ETL has been actively pursuing cloud radar research. The ETL activity in this research field ranges from designing and manufacturing millimeter wavelength radars and microwave radiometers to developing and applying cloud parameter retrieval algorithms. These retrieval algorithms are based on different instrument layouts and vary from simple one-parameter schemes for radar reflectivity measurements only, to the relatively complex multi-sensor methods.

Though most of the ETL algorithms are originally developed for ground-based instrumentation, some of them can be adjusted for use on airborne and spaceborne platforms. This concerns mostly the reflectivity based algorithms and simple multi-sensor methods that are based on radar reflectivity measurements and estimates of cloud optical thickness for ice phase clouds and LWP for liquid water clouds. These complementary estimates could come from the collocated spaceborne radiometer and/or lidar measurements.

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