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Some notes on scattering of radiowaves by clouds

Victor Venema, Jan Erkelens, Herman Russchenberg, and Leo Ligthart
IRCTR, International Research Centre on Telecommunications-transmission and Radar,
Department of Informational Technology and Systems, Delft University of Technology.
Mekelweg 4, 2628 CD Delft, The Netherlands. Fax: +31-(0)15-2784046
Tel: +31-(0)15-2787860, e-mail: V.K.C.Venema@ITS.TUDelft.nl, http://irctr.et.tudelft.nl/~venema/

1. INTRODUCTION

In two recent articles Erkelens *et al.* (1999a and 1999b) showed that coherent particle scatter (scatter of radiowaves by spatial variations in particle mass density) can explain the dual wavelength radar measurements made by Knight and Miller (1998) and dual wavelength measurements of a smoke plume made by Roger and Brown (1997). The 1999-articles also showed quantitatively that coherent particle scatter can be stronger in clouds than incoherent particle scatter. When the spatial variations in humidity and in liquid water are of equal size, coherent particle scatter would be insignificant compared to coherent air scatter – also called clear air or Bragg scatter – (Gossard and Strauch, 1983). The Erkelens *et al.* papers gave some qualitative reasons why the spatial variations in liquid water can be expected to be larger than those in humidity.

The arguments above make a strong case for the significance of coherent particle scatter for radar measurements of clouds. That is why this article will take the first steps in exploring the importance of this scattering mechanism for radar cloud studies. As quantitative data is sparse, these notes must be labelled as tentative, but this is not continuously stressed throughout the article for readability. In the first part (sections 2, 3 and 4) of this article the physical basis of coherent particle scatter will be explained and developed further. In the second part it will be indicated for which radar wavelengths and for which cloud types and atmospheric conditions one can theoretically expect coherent particle scatter to be significant (section 6). Furthermore, it will be shown quantitatively that it is indeed reasonable to assume that coherent particle scatter will dominate coherent air scatter in cumulus clouds. To make this discussion somewhat less academic, this discussion will refer to some cloud measurements in section 5. In the last section some recommendations for further research steps are presented.

2. THEORY OF COHERENT PARTICLE SCATTER

When the N particles in some volume have a completely random distribution in space (also called Poisson distribution), the reflected signal can simply be squared to get the average reflected power:

$$\overline{P}_{inc} = \tfrac{1}{2}\sum_{n=0}^{N} \alpha_n^2 \qquad (1)$$

In this case the average incoherent particle return ($\overline{P}_{inc}$) is only determined by the amplitude (α_n) of the reflected waves, and the phase can be ignored. When for simplicity all particles are assumed to reflect with the same amplitude (α), the average returned power will be $\frac{1}{2}N\alpha^2$, and for small particles (Rayleigh scatter) with number density (ρ) the average radar reflectivity factor will be ρD^6. When there are spatial correlations in the measurement volume, the phase cannot be ignored. In the limit that all amplitudes add up, i.e., when all particles are confined to a volume much smaller than the wavelength, the returned power will be much larger and equals $\frac{1}{2}N^2\alpha^2$.

The spatial correlations in clouds can be thought of in term of voids and lumps of particles at scales from millimetre to kilometre which may be caused by turbulence. In this case the radar backscatter for a volume with droplets will be both coherent and incoherent, and, with one drop diameter (D), is given by Erkelens (1999a):

$$Z \times 10^{-18} = \rho D^6 + 4.2\times 10^{-3} \beta^2 L_0^{-2/3} \rho^2 D^6 \lambda^{11/3} \qquad (2)$$

with ρ the particle number density, L_0 the outer scale length of the inertial subrange of isotropic homogeneous turbulence, and λ the radar wavelength. The first term is the incoherent and the second term is the coherent backscatter, both assuming Rayleigh scatter. The standard deviation of the spatial Liquid Water Content (LWC) variations is assumed to be a fraction (β) of the total LWC. The radar backscatter can be calculated from a volume with a known structure constant of the refractive index (C_n^2). Equation (2) can be derived by two important steps. 1) C_n^2 has to be related to the spatial variations in refractive index. 2) Those refractive index variations can be related to the liquid water variations. The rest of the derivation of equation (2) is straightforward, see Erkelens (1999b).

The radar backscatter is determined by the energy of the three-dimensional power density spectrum ($\phi_n(\mathbf{k})$) in a small spectral band ($\lambda/2 = 2\pi/k$) in the direction ($\hat{\mathbf{k}}$) of the radar beam around half the radar wavelength. Note that also variations at scales larger than $\lambda/2$ contribute, when these variations are not directed parallel to $\hat{\mathbf{k}}$. Following

Ottersten (1969) the spatial power density spectrum integrated over the entire wave-number (k) space is equal to the total spatial variance of the refractive index (var(n)). To compute this three-dimensional power density integral, an assumption has to be made about the shape of the power density spectrum. Equation (3) below is derived with the assumptions that the energy spectrum of the LWC variations follows the well known −5/3 law for homogeneous isotropic turbulence. The largest scale and the smallest scale of the inertial subrange (for which the −5/3 law is valid) is, respectively, designated by L_0 or l_0. If variations at scales larger than L_0 are assumed zero, then using the expressions from Ottersten (1969), the relation becomes:

$$C_n^2 = 5.5(L_0^{2/3} - l_0^{2/3})^{-1} \,\mathrm{var}(n) \approx 5.5 L_0^{-2/3} \,\mathrm{var}(n) \qquad (3)$$

Since scales larger than L_0 contribute to the variance as well, and there is no general theory for this range, the best approach is to only use the part of the refractive index variations that have scales smaller than L_0. In clouds l_0 is in the order of centimetres to millimetres depending on the intensity of the turbulence, which is expected to be much less than L_0, thus the l_0-term is neglected in further calculations.

In the book of Van de Hulst (1981; p. 67 & 70) the refractive index (n) of air with many small spheres is found to be:

$$n = 1 + pKD^3 r / 4 \qquad (4)$$

with $K = (\varepsilon_r - 1)(\varepsilon_r + 2)^{-1}$, a constant that is determined by the refractive index of the particles (ε_r); absorption is neglected. Equation (4) is valid if the second term is small compared to unity, when the size of the particles is small, when the particles are far apart compared to their size and when the particles are close together compared to the radar wavelength. These assumptions are true for a homogeneous cloud and should still be true for the small spatial variations which are considered in this article.

There are two ways to distinguish coherent from incoherent scatter. The first method uses the difference in wavelength dependence of the coherent and incoherent backscatter. The incoherently reflected power from a droplet is strongly dependent on the wavelength, whereas the coherent backscatter from turbulent variations is less wavelength dependent. The radar reflectivity factor corrects for the incoherent wavelength dependence, thus the radar reflectivity factor of an incoherently scattering volume is the same when measured by two radars with different wavelengths. For a coherently scattering volume the difference in radar reflectivity measured by two radars differs by a factor. For a given slope of the turbulence energy spectrum and the ratio of the two wavelengths, this factor can be calculated. For example, the difference in radar reflectivity factor between a 10-cm and a 3-cm radar is 19 dB, when the slope of the spectrum is −5/3. Details can be found in, e.g., Knight and Miller (1998).

Another method is to look at the angular dependence of the scatter using a bi-static radar, see for details Gossard and Strauch (1980). Incoherent scatter does not depend on the azimuth for vertically polarized radiowaves, whereas coherent scatter does change with the azimuth.

3. MEASURED SPATIAL VARIATIONS

There is a large set of in-situ measurements of spatial variations in literature; nowadays at very small scales. Some of these measurements will be summarized here. For coherent scatter *physically* significant spatial variations in liquid water content and humidity are of interest. Often, however, just measurements of spatial variations in number density are tested for *statistical* significant deviations from the Poisson statistics. A volume with statistical significant deviations may not give significant coherent scatter.

Davis *et al.* (1999) have made measurements of spatial LWC variations down to scales of 4 cm with a Particulate Volume Monitor (PVM-100A) probe in broken-stratocumulus clouds with embedded towering cumulus clouds during the Southern Oceanic Cloud EXperiment (SOCEX). The average LWC in the cloud is 0.290 ± 0.167 g/m^{-3}, thus the relative standard deviation is 58 percent. They find a significant change of the slope (a scale break) at 2 to 5 m. At longer scales the slope of the variance spectrum is close to −5/3 (1.6 ± 0.1), but at smaller scales there are more variations, the slope here is -0.94 ± 0.10. These extra variations correlate with the occurrence of spikes in the LWC (voids or blobs at most a few 4-cm pixels wide). Davis *et al.* hold these spikes responsible for the extra spatial variations, below 2 to 5 m.

In an earlier article with LWC measurements of stratocumulus Davis *et al.* (1996) also find spikes, which in some parts are above the average and in others well below the average. Other parts are relatively smooth. The average relative standard deviation was about 19 %, but it varied highly per measurement; the lowest value found was 5 %.

One of the first measurements of cm-scale spatial variations was made by Baker (1992). Measurements of cumulus clouds at all heights were made with a Forward Scattering Spectroscope Probe (FSSP) with a spatial resolution of about 0.32 mm. To the measured number of droplets he applied a test on the variance relative to the mean. Baker found statistically significant spatial correlations in number density on cm-scales in the majority of cloud penetrations. However, statistically significant variations on all measured scales were only observed in small parts (often near edges) of many clouds and throughout a few clouds.

Malinowski *et al.* (1994) observed that entrained air contained filaments of cloudy air with droplet concentrations close to those observed in the clouds. Furthermore,

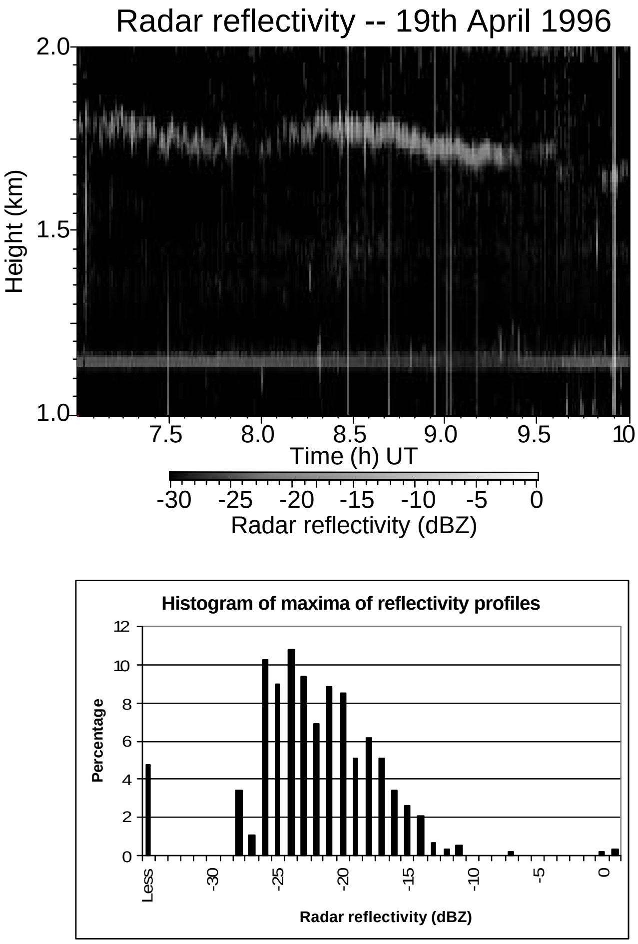

Figure 1. Measurement of stratocumulus cloud (1a) on the 19th of April 1996 by the Delft Atmospheric research radar, a 9 cm FM-CW radar. The histogram (1b) shows the maximum values of the radar reflectivity profiles between 8 and 9:30 hrs UT of the stratocumulus cloud shown in fig. 1a.

they suggest that the distribution of these filaments is anisotropic.

Jameson *et al.* (1998) made measurements using Particle Measurement System (PMS) two-dimensional optical array probes with a resolution of 130 m in a tropical warm precipitating cumulus about 1 km above the cloud base. The main conclusion according to the authors is that spatial variations in drop counts are significant from 100 m up to 2 km scales (variance is much larger than the mean). Besides that they also found variations down to 5 cm scales, using the distribution of the interdrop distance.

Korolev and Mazin (1993) have made extensive measurements – 50 cloud layers on 20 different days with a total length of 1710 km – with an FSSP-100 of stratiform clouds (stratus, stratocumulus, altostratus, altocumulus, and nimbostratus). Based on this dataset they conclude that cloud holes occur most frequently in the vicinity of the upper and lower boundaries, but also in parts separated from the upper and lower boundary by hundreds of meters. About 80 percent of these holes was of the smallest size (up to 10 m). On average the holes (defined as volumes with less than 50 percent of the average number density) had a volume of about 7 percent. However, this number varies highly: in some cloud this was 20 percent and sometimes no holes were found for dozens of kilometers. Also regions with increased droplet number concentration were found. However, these were mainly due to the appearance of a large number of small droplets, so these may not be that important for the spatial variations in LWC.

Kozikowska *et al.* (1984) made a hologram of 22.5 cm^{-3} in fog to measure the 3 dimensional droplet distribution on the smallest scales. The distribution of the number density in this one sample taken was significantly not Poisson.

Humidity variations have been measured by Politovich and Cooper (1988). They estimate the supersaturation in cumulus clouds with a resolution of 10 m by measuring the vertical velocity (Rosemount 858 gust probe) and the drop size distribution (FSSP). The supersaturation was estimated to be in the range of -0.5 to 0.5 % for all cloud regions during 147 cloud penetrations of 13 clouds on 8 days. In the entrained regions the standard deviation was below 0.4 % and in the core of the cloud around 0.1 %.

Concluding, measurements of cumulus, stratus clouds and fog, with a variety of instruments indicate that non-Poisson distributed droplets occur regularly. However, Poisson distributions are found as well. Especially, the presence of spatial variation in the adiabatic cores of cumulus clouds is still under debate, see Grabowski and Vaillancourt (1999) and references therein. The structure of the variations is described as spikes (Davis, 1996) and anisotropic filaments (Malinowski *et al.*, 1994); the theory of coherent particle scatter may have to be modified to account for such strong variations.

Unfortunately quantitative measurements of spatial LWC variations could only be found for stratocumulus (with embedded cumulus) and for some cloud types no measurements of spatial distribution could be found at all. Only one indirect measurement of spatial humidity variations was found in literature. Unfortunately no simultaneous measurement of humidity and LWC variations were found in literature. The strength of the coherent backscatter can only be estimated for the few quantitative measurements, which will be done in section 6.

4. SOURCES AND SINKS OF SPATIAL VARIATIONS

To be able to extrapolate the sparse measurement data to other cloud types and atmospheric conditions, the sources and sinks of these variations have to be known. There are a number of possible sources for spatial variations of LWC, the first three are taken from the study of Korolev and Mazin (1993) on stratiform clouds.

Entrainment of dry air from outside the cloud is especially important at the cloud top, amongst others due to loss of stability of air parcels by radiative cooling. Entrainment or mixing can also be important at the sides of cumulus clouds. The spatial humidity variations caused by mixing, give rise to mantle echoes at the cloud edges on dry days. These mantles can be up to 1 km thick, see e.g. Knight and Miller (1998).

At the base of the cloud variations can be created by differences in the condensation level of the air parcels, due to differences in the temperature and humidity of the initial air parcels. Korolev and Mazin (1993) estimate that this can cause variations in the condensation level in the order of tens of meters. This fits with the observation of a relatively higher number of cloud holes in the lowest 100 m of stratiform clouds compared to the middle of the cloud.

Spatial variations can also be caused by vertical movements in a sub-adiabatic cloud. For a typical stratiform cloud, Korolev and Mazin estimate that a descending movement of 60 m can create a cloud hole.

Spatial LWC variations may also be created by the inertia of the droplets in a turbulent field, see e.g. Squires and Eaton (1991), Shaw *et al.* (1998), Vaillancourt (1998) and Pinsky *et al.* (1999). A particle in a vortex will experience a centrifugal force outward, thus after some time a vortex can become particle free and the particles will be collected in the quiet areas with low vorticity. It is still under discussion whether the right conditions are present in (cumulus) clouds, as modeling of realistic turbulence for clouds is still too difficult. Important unknown factors are the lifetime of the vortices, the volume fraction of the vortices and the influence of sedimentation due to gravity. If true it would be an important mechanism as it can also act in the adiabatic cores of clouds.

When making statements about (the absence of) mixing one has to remember that an adiabatically ascending air parcel is an idealization. A parcel always has some heat exchange, whether this is a significant deviation from adiabicity will depend on the application. For example, the S-band reflectivity in fig. 6 in the paper of Knight and Miller (1998) shows significant spatial variations in humidity or LWC in a 500 m thick mantle echo, which are thought to be caused by *mixing* with environmental air. Although the X-band reflectivity shows flat echo bases at already a few tens of meters from the cloud edge. These echo bases are interpreted by Knight and Miller as *unmixed* ascent of the cloud air parcel. These two statements do no contradict each other since the amount of mixing to create coherent scatter is probably less than the mixing needed to a get significant decrease in LWC.

Sinks for LWC variations at large scales are turbulence (which transfers it to smaller scales) and for LWC variation at small scales a sink is sedimentation. The thermal diffusivity of droplets is very small, but maybe turbulence can also be a sink for variations on scales smaller than the dissipation range due to the inertia of the droplets.

Concluding, all the above mechanisms – entrainment, parcel differences, vertical movements in sub-adiabatic cloud and inertia of droplets in turbulent medium – depend on the strength of the turbulence. The first three effects will become stronger in a more turbulent situation. For the effect of the inertia of drops the relation may be more complex. As mixing with environmental air should be an important mechanism at cloud edges and the gradients are high at the top and sides of the clouds, the spatial variations are expected to be biggest at these cloud edges. For stratus clouds this is confirmed by Korolev and Mazin (1993) who state that cloud holes are found more often at cloud tops. For cumulus this is confirmed by the mantle echoes.

Not much is known about the sources and sinks of humidity variations in clouds. Sources should be entrainment, the finite relaxation time of evaporation and condensation after the temperature of the parcel changed due to turbulence of an up or downdraft. Sinks should be turbulence, diffusion and evaporation and condensation. Thus also humidity variations are expected to be largest at the cloud boundaries.

5. RADAR MEASUREMENTS

This section gives some examples of radar cloud measurements in which coherent particle scatter may play a role, in order to show in which cases one may expect coherent particle scatter. The three examples are: young cumulus clouds, winter clouds and a stratocumulus cloud. There are also strong indications of coherent particle scatter in a dual-wavelength measurement of a smoke plume, Erkelens *et al.* (1999b).

In a recent article, Knight and Miller (1998) discuss a large number of measurements of developing cumulus clouds, performed with 2 radars: an X-band radar and an S-band radar with wavelengths of 3 and 10 cm, respectively. Most measurements could be explained by the traditional theory. However, on humid days, the patterns for both radars looked similar, resulting in a correlation between the S- and the X-band reflectivity factors. This correlation is puzzling in cases where the difference in reflectivity factors is not equal to 0 dBZ (for purely incoherent scattering) or 19 dBZ (for purely coherent scattering). The X- and S-band reflectivity factors in some measurements lie on a line with a slope of one, with a typical difference of about 10 dBZ; other offsets have been measured as well. The phenomenon is mostly confined to the core of the reflectivity pattern and its nearby surroundings and to the region near the cloud base.

A similar correlation was measured by Baker *et al.* (1998), who already speculate that the droplets may scatter

Cloud type	N_d (cm^{-3})	Comment	Source
water clouds			
Cumulus	800-900	At cloud base of similar clouds as Knight and Miller in Florida.	Paluch *et al.* 1996
Continental Cumulus	500-800	30 minutes of data in Montana.	Politovich and Cooper, 1998
Continental stratocumulus	200-500	One measurement.	Korolev and Mazin, 1992
Continental stratus	347	about 1.5 hrs data.	Sassen *et al.* 1999
Coastal stratus	75-150	One measurement.	Korolev and Mazin, 1992
Fog	50	One volume of 22.5 cm^{-3}.	Kozikowska *et al.*, 1984
Fog	1 to few hundred		Pruppacher and Klett, 1997
ice clouds			
Altostratus	Up to 10		Pruppacher and Klett, 1997
Alto stratus Alto cumulus	75	One measurement.	Korolev and Mazin, 1992
Cirrus	0.05 to 0.5		Pruppacher and Klett, 1997
Cirrus	Up to 0.2	Model, highest N_d at top.	Sassen and Khvorostyanov, 1998
Ice fog	100-200		Pruppacher and Klett, 1997

Table 1. Number density found in clouds.

coherently. A quantitative explanation for these correlations in terms of coherent particle scatter is given by Erkelens *et al.* (1999a). Knight and Miller also observed cases in which the difference between the reflectivity factors was more than 19 dB (up to 22 dB) in the mantle echo. This could indicate that the slope of the humidity spectrum is steeper than −5/3 (−1.67). In the case of 22dB difference the slope would be −2.2.

Using a setup with two X-band radars Gossard and Strauch (1980) were able to separate coherent from incoherent scatter and furthermore to measure the coherent scatter as a function of wave number. During clear air situations they found a slope of the energy spectrum of −1.7, near −5/3. However, during cloudy situation the slope was less steep: −0.9.

A measurement during CLARA (The Dutch CLouds And RAdiation measurement campaigns; Van Lammeren *et al.*, 1999) of a stratocumulus cloud is shown in fig. 1a. The radar reflectivity measured by the 9-cm wavelength Delft Atmospheric Research Radar (DARR) is high. A histogram of the maximum values of the reflectivity profiles is shown in fig. 1b. Values above −20 dBZ are not uncommon, whereas in-situ drop size distribution measurements with an FSSP-100 never give values above −25 dBZ for incoherent droplet scatter. This suggests that the scatter may be enhanced by coherent scattering. One can however not be conclusive as a few large particles can enhance the radar reflectivity a lot, whereas the FSSP may miss these particles due to the small sampling volume and the small maximum size. The airplane also did not fly at the height of the maximum reflectivity values, during the time the highest values occurred. The difference in radar reflectivity values is intriguing though.

6. DISCUSSION OF THE MEASUREMENTS

In this section the radar measurements from the previous section will be discussed in terms of coherent particle scatter, coherent air scatter and incoherent particle scatter, using the in-situ measurements as a reference for the spatial variations.

6.1 Coherent air scatter

Using the measurement of spatial humidity variations by Politovich and Cooper (1988) it will be shown that coherent particle scatter can also dominate coherent air scatter for cumulus clouds. Gossard and Strauch (1983) calculated that if the spatial variance of the water vapor content is equal to the spatial variance of liquid water content, the radar reflectivity of the humidity variations should be about 28 times larger. In this calculation the slopes of both energy density spectra where assumed to be equal. In other words, for equal reflectivity factors the spatial standard deviation of the liquid water content should be 5.3 times larger than the spatial standard deviation of the humidity variation. In the article by Knight and Miller (1998) the regions giving a high correlation between S and X-band are about 2 km above cloud base. At this height roughly half of the water vapor will be condensed into liquid water in an adiabatic cloud with a temperature at cloud base between 20 and 30 centigrade. This cloud base temperature is taken from Paluch *et al.* (1996) who measured similar clouds as Knight and Miller.

The relative standard deviations in the spatial humidity variations measured by Politovich and Cooper (1988) were: 0.4 % in 80 percent entrained air and 0.1 % in the core of cumulus clouds. For the coherent particle scatter to dominate coherent air scatter the spatial standard deviations of

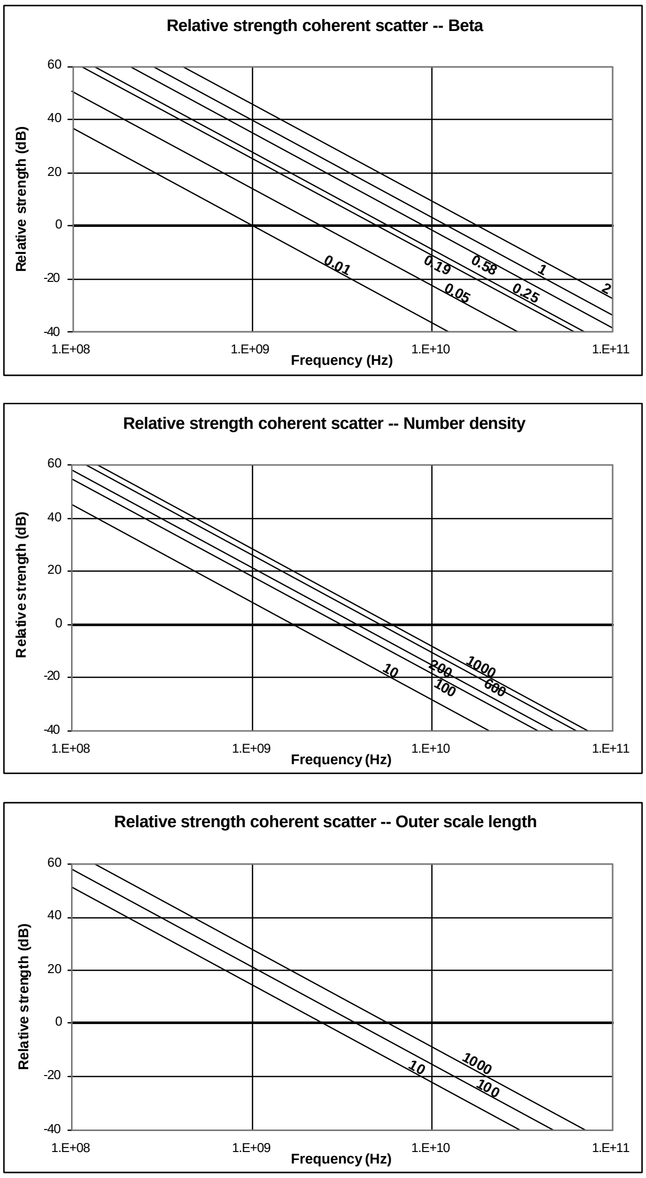

Figure 2. Plots of the relative strength of coherent particle scatter compared to incoherent particle scatter as a function of radar frequency as calculated by eq. (2). Fig. 2a shows the influence of the relative standard deviation of the LWC. Fig. 2b shows the influence of particle number density and fig 2c shows the relation for some values of the outer scale of the inertial subrange L0. The assumptions for the values of the other variables for each plot are in section 6.2.

the LWC variations should be 5.3 times bigger, thus above 2 % for the entrained region and about 0.5 % in the core of cumulus clouds. Lower in the cloud the ratio of water vapor and liquid water will be higher, thus coherent *air* scatter will be more important near cloud base. To explain the dual-wavelength radar measurements of Knight and Miller (1998), Erkelens *et al.* (1999a and 1999b) needed about 25 % LWC variations. Given the low standard deviations in the LWC variations needed to dominate coherent air scatter which were computed above and the high values measured (5 to 58 % in stratocumulus; Davis *et al.*, 1999 and 1996), the scatter by the water droplets should dominate coherent air scatter in cumulus clouds, except close to the cloud base. The assumption made then is that the LWC variations in cumulus are in the same order as those in stratocumulus.

In stratocumulus the LWC is much lower. When the standard deviation in LWC is 0.167 g/m^{-3} Davis (1999), the standard deviation in humidity should be 0.03 g/m^{-3} for the coherent scattering to be of equal size. This corresponds to a humidity of 7.5 g/m^{-3} with 0.4 % standard deviation or to a humidity of 30 g/m^{-3} with 0.1 % standard deviation. As these humidity values are of a natural order of magnitude, a conclusion for these clouds cannot be drawn yet.

6.2 Incoherent particle scatter

To see how important incoherent scatter is compared to coherent particle scatter, one has to compare the two terms in eq. (2). In Erkelens *et al.* (1999a) it was already shown that coherent particle scatter can dominate incoherent particle scatter for the 10-cm wavelength radar reflection of the cumulus cloud by Knight and Miller (1998) using values from literature. The coherent term is about 10 dB stronger using the values: $L_0 = 10$ m (from Van Zandt *et al.*, 1978), $\rho = 800 \cdot 10^6$ m^{-3} (from Paluch *et al*., 1996), $\beta = 0.25$, $\lambda = 0.1$ m.

To investigate this relation for other situations the influence of the variables in eq. (2) is plotted in a few graphs. In fig. 2a the relative strength of coherent particle scatter compared to incoherent particle scatter is plotted for different values of the relative spatial standard deviation of LWC (β) taking the other values for a cumulus case: $L_0 = 10$ m, $\rho = 900 \cdot 10^6$ m^{-3}. The values for β are taken from the previous discussion, including some more extreme examples. The high values may occur at cloud boundaries.

In fig. 2b the relative strength of coherent particle scatter compared to incoherent particle scatter is investigated in relation to the number density (ρ) taking for the other values: $L_0 = 10$ m, $\beta = 0.25$. Measured number densities from literature are summarized in table 1; there is a considerable spread in the naturally occurring number densities. A number density of 1000 cm^{-3} is a high value for cumulus, 200 to 500 cm^{-3} is representative for continental stratus clouds, the value of 10 cm^{-3} corresponds to some ice clouds.

That the outer length scale is not very important can be seen in fig. 2c. The other values are taken to be: $\rho = 900 \cdot 10^6$ m^{-3}, $\beta = 0.25$. The right value for the outer scale is uncertain, and it will be related to the value for β. This plot with L_0 as independent variable is just included to show that the spread is not so large.

Concluding, for X-band radar coherent particle scatter can dominate for the highest β and number density. For mm-wave radars coherent particle scatter is not likely. For windprofilers coherent particle scatter will normally dominate incoherent particle scatter. For typical number densities for stratus clouds, coherent particle scatter may dominate incoherent scatter for cm-radars. For typical number densities for ice clouds one will need a windprofiler for coherent scatter to dominate. Given the poor present state of knowledge the above conclusion may be modified in the future.

6.3 Slope of energy spectrum

Assuming fully-developed isotropic turbulence one gets a slope of −5/3 in the inertial subrange when the sources of the variation act on scales larger than the outer scale (L_0) and the sinks on scales smaller than the inner scale (l_0). For other turbulent conditions there is no theory available. In all computations in section 6.2 the slope of the humidity and LWC spectra was taken to be −5/3, which may not be true for humidity and LWC. For a passive *conservative* additive the slope of the spectrum will be the same as the slope of the turbulent energy spectrum, Tatarski (1961). However, both the amount of water vapor and liquid water are not conservative in clouds. One can expect the slope to become more flat (steep) if there is an additional source (sink) of variations within the inertial subrange. A sink for humidity variations within the inertial subrange could be evaporation and condensation. The effect of droplet inertia and entrainment could act as a source for LWC variations above the inner scale.

The difference in radar reflectivity in a dual-frequency measurement of a purely coherently scattering volume is a function of this slope. A steeper slope will give a larger difference and vice versa.

Knight and Miller (1998) found that the reflectivity factor difference between 10 and 3 cm is higher than 19 dB in the mantle echo, which is an indication that the slope of the humidity variations is steeper than −5/3. Davis *et al.* (1999) found a flatter slope in the LWC variations at scales below 5 m. Gossard and Strauch (1981) found a flatter slope than −5/3 in the coherent scatter of clouds of a not specified type. It is not clear if this coherent scatter was from humidity or LWC variations.

Deviations in slope from −5/3 can be important and if large enough such a deviation may change some conclusions from this article. If the slope of LWC would be more flat, coherent particle scatter would become more important for small wavelengths. If furthermore the slope of the humidity variations would be steeper, the relative strength of coherent particle scatter compared to coherent air scatter would become larger at small wavelengths. Thus for radar measurements of coherent scatter it is important that the slope of the energy spectrum is measured – preferably of humidity and LWC at the same time – and that the underlying mechanisms that influence the slope are well understood.

7. SUMMARY AND CONCLUSIONS

This paper explored the possibility of significant coherent particle scatter in clouds. Measurements in literature have shown significant spatial variations in stratus, stratocumulus and cumulus clouds. Quantitative measurements of spatial LWC variations are unfortunately only available for stratocumulus with embedded cumulus and quantitative spatial humidity measurements only for cumulus clouds.

Dual-wavelength radar measurements of cumulus clouds show signs of coherent particle scatter. Furthermore there is evidence that the slope of the humidity and LWC spectra can deviate from the theoretical −5/3. And 9-cm wavelength radar measurements of stratocumulus might give more reflections than can be explained by incoherent particle scatter.

Preliminary theoretical calculations show that coherent particle scatter can dominate both coherent air scatter and incoherent particle scatter in some cases. Given a slope of −5/3 incoherent particle scatter should dominate coherent particle scatter for mm-wave radar, for cm-wave radar it will depend on the cloud type and atmospheric conditions. For dm-wave radar coherent particle scatter is almost always important compared to incoherent scatter. For S-band radar coherent particle scatter of stratus and stratocumulus clouds could be significant, however there is no reliable estimate of the relative strength of coherent air scatter for these types of clouds.

8. REMAINING QUESTIONS

Since not much attention has been given to coherent particle scatter until now, there are still many remaining questions. To test the theory of coherent scatter, simultaneous measurements of LWC and humidity variations should be compared to radar measurements of coherent scatter. Experiments on cumulus similar to Knight and Miller but with mm-wave radar as smallest wavelength or a bi-static radar would give more confidence when the results would be similar.

To further develop the theory small scale in-situ measurements should be made of humidity, LWC and temperature, preferably in the same volume, so that also the covariances at small scales can be determined. Measurements of the LWC and humidity spectra should be made for a range of different cloud types, to determine for which cloud types coherent scatter is important.

Radar measurements could make a contribution to the research on LWC and humidity variations, which are important for the understanding of the broadening of the drop size distribution and consequently for warm rain formation. In combination with other remote sensing instruments the number density may be determined, Erkelens *et al.* (1999c). A dual-wavelength radar can then estimate the magnitude of the spatial variations.

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