Atmos. Chem. Phys. Discuss., 14, 11153–11176, 2014
www.atmos-chem-phys-discuss.net/14/11153/2014/
doi:10.5194/acpd-14-11153-2014

Atmospheric Chemistry and Physics
Open Access
Discussions

The relative dispersion of cloud droplets: its robustness with respect to key cloud properties

E. Tas[1,2], A. Teller[1], O. Altaratz[1], D. Axisa[3], R. Bruintjes[3], Z. Levin[4,5], and I. Koren[1]

[1]Department of Earth and Planetary Sciences, Weizmann Institute of Science, Rehovot, Israel
[2]The Department of Soil and Water Sciences, The Robert H. Smith Faculty of Agriculture, Food and Environment, Hebrew University of Jerusalem, Rehovot, Israel
[3]Research Applications Laboratory, National Center for Atmospheric Research, Boulder, CO, USA
[4]Department of Geophysics and Planetary Science, Tel Aviv University, Israel
[5]The Energy, Environment and Water Research Center, the Cyprus Institute, Nicosia, Cyprus

Received: 25 March 2014 – Accepted: 8 April 2014 – Published: 6 May 2014

Correspondence to: I. Koren (ilan.koren@weizmann.ac.il)

Published by Copernicus Publications on behalf of the European Geosciences Union.

Abstract

The relative dispersion (ε) of cloud droplet size distribution, defined as the ratio between cloud droplet size distribution width (σ) and cloud droplet average radius ($\langle r \rangle$), is investigated using airborne measurements of warm cumulus clouds. The data is used to study the relation of ε with microphysical and thermodynamic characteristics of the clouds. The results show that ε is constrained with average values in the range of ~ 0.25–0.35. It is shown that ε is not correlated with the cloud droplet concentration or with the Liquid Water Content (LWC). However, the relative dispersion variance (related to the third moment of the droplets distribution) shows sensitivity to the droplets' concentration and LWC, suggesting smaller ε variability in more adiabatic regions in the clouds. A clear criterion for the usage of the in situ airborne measurements data for statistical moments' calculations is suggested.

1 Introduction

The droplet size distribution is one of the key characteristic properties of clouds. The size distribution properties are controlled by the microphysical and dynamical state of the cloud. Near cloud base at the first stages of droplets formation the size distribution is determined by the supersaturation (determined by the thermodynamic conditions and the updraft) and by the aerosol properties. At later stages of the cloud development additional processes like collision-coalescence, rain drops sedimentation, drag-forces effects, entrainment and mixing further modify the size distribution. Even on later stages of the cloud development the droplet size distribution affects the efficiency and magnitude of the different microphysical processes and their effects on the dynamics through drag of the falling drops with respect to the updraft, terminal velocities and the release of latent heat.

The relative dispersion of cloud droplets (ε) is a parameter that represents the droplets size distribution. It is defined as the ratio between the standard deviation (σ)

ACPD
14, 11153–11176, 2014

and the mean radius ($\langle r \rangle$) of the distribution. Both σ and $\langle r \rangle$ are key variables as they are used in various model schemes, like for determining the reflectivity of clouds (Hansen and Travis, 1974; Slingo, 1989; Liu and Daum, 2000a, b; Daum and Liu, 2003) and for calculating autoconversion processes (e.g., Liu et al., 2005, 2006a; Hsieh et al., 2009). Often, instead of using both σ and $\langle r \rangle$, their ratio (ε) is used. It is done in atmospheric models spanning various scales from cloud resolving models (CRM's) to global climate models (GCM's). These models are used to explore aerosol effects on clouds, such as the first aerosol indirect effect (linking higher aerosol loading to the formation of numerous but smaller cloud droplets and higher clouds' reflectivity (Twomey, 1977)), the second indirect effect (linking the increase in aerosol loading with an increase in cloud lifetime (Albrecht, 1989)), and convective cloud invigoration (Koren et al., 2005; Andreae et al., 2004; Tao et al., 2012; Altaratz et al., 2014).

Considering the importance of ε, many studies have been conducted aimed at analyzing the sensitivity of this parameter to the environmental conditions and to key microphysics and thermodynamic cloud properties. The reported results are quite diverse. Many studies examined the impact of various physical factors on ε in startiform clouds (Peng and Lohmann, 2003; Rotstayn and Liu, 2003; Peng et al., 2007; Miles et al., 2000; Martin et al., 1994; Ma et al., 2010; Lu and Seinfeld, 2006; Pawlowska et al., 2006). Fewer studies examined it for convective clouds and some of them are presented below. Lu et al. (2008) and Berg et al. (2011) analyzed airborne measurements of shallow cumuli and found an average ε of around 0.3. Zhao et al. (2006) analyzed data collected in 135 flights in different environments and found ε values in the range of 0.4 to 0.5, when the droplet concentration (N_c) was higher than 50 cm^{-3}. Similar results were reported by Deng et al. (2009). Martins and Silva Dias (2009) studied cumulus clouds in the Amazonian dry season and found ε values in the range from 0.38 to 0.59.

In some of the studies the sensitivity of ε to the aerosol loading was examined. By compiling measured data from different field studies, including warm cumulus (Cu), Liu and Daum (2002) suggested that ε is sensitive and positively correlated with aerosol

Discussion Paper | Discussion Paper | Discussion Paper | Discussion Paper |
ACPD
14, 11153–11176, 2014
The relative dispersion of cloud droplets
E. Tas et al.

loading at cloud base. Subsequently, this conclusion was supported by modeling (Yum and Hudson, 2005), observational data (McFarquhar and Heymsfield, 2001) and by theoretical studies (Liu et al., 2006b). Other investigations using observational data (Martins and Silva Dias, 2009; Hsieh et al., 2009) suggested a negative relationship between ε and the aerosol loading.

ε is influenced by other factors as well. Yum and Hudson, (2005) and Liu et al. (2006b) studied the combined effect of updraft values and aerosol loading (as they both determine the supersaturation values) on ε using adiabatic growth theory of cloud droplets. These factors are influential mainly at early stages of cloud development before collection, sedimentation, entrainment and mixing become dominant. The authors found an opposite effect: an increase in ε with the increase in aerosol loading and a decrease in ε with an increase in the updraft velocity. They claimed that more continental clouds have smaller $\langle r \rangle$ and, due to that, larger ε. The response of σ was different between the model calculated values and observations. Namely, larger in their model results and smaller in observations for continental clouds (Yum and Hudson, 2005). Weaker updrafts were associated with larger σ, and larger $\langle r \rangle$, resulting in lower cloud supersaturations.

The chemical composition of the aerosols is another influencing factor that should be taken into account when studying ε. It determines the activation process and the growth by condensation and therefore affects the spectral distribution of the drops and the behavior of ε (Martins and Silva Dias, 2009).

Zhao et al. (2006) and Liu et al. (2008) indicated that the so-called dispersion effect (higher ε in high aerosol loading environment) may account at least partially for the discrepancies between different estimations of the first indirect effect in different studies (Feingold et al., 2003; Rosenfeld and Feingold, 2003). This effect is linked to the impact of ε on the calculated effective radius. The climate models that quantify the first indirect effect must account for the influence of the cloud droplet spectrum on the effective radius, and in cases that ε is positively correlated with the aerosol loading it can result in larger effective radius and weaker cloud reflectivity (Slingo, 1990).

Xie et al. (2013) studied four types of parameterizations for treating the relationship of the droplet concentration and ε. They implemented these schemes into the Weather Research and Forecasting (WRF) model, aiming at studying the effects of aerosol on cloud microphysics and ground precipitation. They concluded that the Nc-ε relationship (positive or negative change of ε with Nc) influences the autoconversion processes from cloud droplets to raindrops and therefore affect the response of ground precipitation to the change in aerosols. This paper suggested that for positive Nc-ε relationship the large-sized rain drops at high aerosol concentrations enhance the efficiency of the autoconversion process and the surface precipitation. The diversity of schemes for the relationship of Nc-ε (as shown in Fig. 1 of Xie et al., 2013) suggests that much more research is required to understand the physics behind the properties of ε.

Recently, Tas et al. (2012) suggested a new way to monitor the response of ε in warm Cu clouds to the ambient conditions. In that work the cloud lifetime was divided into three stages based on the microphysical evolution of the cloud. Using a detailed microphysical model, it was shown that the microphysical evolution of the cloud imposes a most significant impact on ε, while aerosol loading and dynamics play only a secondary role. Their results indicated that ε has a narrow range of around ~ 0.25–0.35, during the mature stage of clouds (defined as the stage when the total water mass is around its maximum and is relatively stable). They claimed that trends in ε can be explained by the balance between the two main growth processes that dictate the droplet size distribution, condensation and collision-coalesce (before the initiation of significant rain). At the mature stage the relative importance of the collision-coalescence induced growth slowly increases, such that ε growth is relatively slow.

In this study we use detailed airborne cloud measurements carried out in 2007–2008 as part of the Cloud and Aerosol Research in Istanbul (CARI) field campaign (Teller et al., 2008). The aim of this campaign was to study the characteristics of the clouds' droplet size distributions and the relation between them and other cloud properties at different vertical levels in Cu clouds deeper than 1 km. For this study we examined non-precipitating clouds. The convective cloud properties were measured under dif-

ACPD
14, 11153–11176, 2014

ferent aerosol loadings, examining both the (more adiabatic) core of the cloud and its boundaries that are influenced more significantly by entrainment.

2 Measurements and instrumentation

The 2007–2008 Cloud and Aerosol Research in Istanbul (CARI) was aimed at studying cloud and precipitation characteristics as a feasibility study for cloud seeding operations in the area of Istanbul (Teller et al., 2008). A Piper Cheyenne II research aircraft (see Axisa et al., 2005 for details of the aircraft) was equipped with a Droplet Measurement Technologies (DMT) Cloud Droplet Probe (CDP) for measuring the concentration and size distribution of cloud droplets in the radius range of 1.5–25 µm. In addition, aerosol concentrations and size distributions in the radius range of 0.055–1.5 µm were measured using a DMT Passive Cavity Aerosol Spectrometer Probe (PCASP, SPP200). This work is focused on the measurements that were carried out on the 6–7 June 2008.

A frontal system passed over the area of Istanbul on the night of 6 June 2008 bringing some rain showers to the area. The airborne measurements in 5 warm cumulus clouds were conducted before and after the passage of the front. The upper panel of Fig. 1 shows a Moderate Resolution Imaging Spectroradiometer (MODIS) image that was taken on 7 June 2008 above the Eastern Mediterranean region, showing the cloud cover west of Istanbul, where the airborne measurements were carried out.

The exact time of each cloud measurement case, its abbreviation and the measured aerosol loading are given in Table 1.

Flight of 6 June 2008 – on this day the research aircraft conducted cloud penetrations in two pre-frontal convective clouds that developed west of the urban area of Istanbul. These clouds are hereafter referred to as TRK1 and TRK2 (see Fig. 1). The thickness of the two clouds was around 2000 m with cloud base temperature of about 15 °C and cloud top temperature between 0 and −2 °C. The warm cloud top temperature suggests that the clouds did not contain ice hydrometeors.

ACPD
14, 11153–11176, 2014

Flight of 7 June 2008 – on this day the research aircraft conducted measurements in three post-frontal cumulus clouds that developed west of the urban area of Istanbul. These clouds are referred to as TRK3, TRK4 and TRK5 (see Fig. 1). The environmental conditions of clouds TRK3 and TRK4 and their physical sizes are quite similar to TRK1 and TRK2. TRK5 was a shallower cloud of only 1000 m depth and cloud top temperature of 8 °C.

To assure statistically significant results we analyzed cloud measurements with N_c larger than 10 cm^{-3}, in agreement with the analysis and cloud definition of Deng et al. (2009). Sensitivity tests revealed that the results did not change significantly by applying other threshold values for defining the clouds' boundaries, within the range 5 and 50 cm^{-3}. Analyses were done for two types of regions in the clouds: (1) inner-cloud and (2) cloud-boundary.

The more adiabatic, in-cloud dataset, required that not only the sampling point itself, but also two neighboring sampling points (representing in total 2 s or ~ 140 m) are all associated with N_c higher than 10 cm^{-3}. The cloud-boundaries were defined as those data intervals that met the condition of $N_c > 10$ cm^{-3} but one of its neighboring points did not meet this criterion.

3 Results

Figure 1b shows some differences between the clouds that were investigated on 6 and 7 June. The measurements of TRK1 and TRK2 (6 June 2008) show an increase and then a decrease in the maximum of the total cloud droplet concentration as a function of altitude (from a maximum of 1650 cm^{-3} at cloud base to 1100 cm^{-3} at cloud top in cloud TRK1 and from 1400 to 700 cm^{-3} in cloud TRK2). The ambient aerosol concentration in the diameter range of 0.11–3 µm below cloud base was about 1800 cm^{-3} in cloud TRK1 and 900 cm^{-3} in cloud TRK2.

The cloud droplet concentrations that were measured on 7 June 2008 were smaller than those measured during 6 June, and in the majority of the penetrations were

ACPD
14, 11153–11176, 2014

smaller than 1000 cm^{-3} (see Figs. 1 and 2). The average aerosol concentration in the morning flight (TRK3) was 700 cm^{-3} while in the afternoon flights (TRK4 and TRK5) it increased to 1000 cm^{-3} as a result of regeneration of the air pollution layer in the vicinity of Istanbul after the cleaning by the rain the night before.

Figure 2 shows the average droplet size distribution as a function of height. The height above ground level is binned into 10 sectors and is presented as a function of the corresponding mean droplet radius in each sector. The mean droplet radius ($\langle r \rangle$) and the standard deviation (σ) at each height bin is shown by the yellow and the green lines, respectively. Note that due to the limitation of the measured drops size by the cloud drop probe, CDP, the spectrum of the distribution is truncated at a radius of $\sim 20\,\mu m$.

Such representation of the size distribution allowed us to investigate the impact of the instrument limitation on our analysis. It revealed that in all cases except TRK3 the upper limit cut-off droplet size was below 20 µm, implying that no droplets > 20 µm were present. TRK3, having a lower aerosol loading (see Table 1) had a lower cloud droplet concentration (see Fig. 3) and a broader size distribution. Thus, although a contribution from larger droplets is expected, it could not be included in the analysis. Such a limitation suggests that calculations of moments of the size distribution might be biased in the case of TRK3.

Figure 2 reveals additional information about the sampled clouds: (i) Cases TRK1 and TRK2 had larger droplet concentrations compared to the other clouds. In these cases the $\langle r \rangle$ was $\sim$ 5–6 µm at cloud base and between 8–9 µm near the cloud top. As case TRK3 had the lower droplet concentration, the average radius at cloud base was 6.5 µm and increased to about 8.5 µm close to the cloud top. (ii) In cases TRK4 and TRK5 the width of the droplet size distributions decreased with altitude in the clouds. In general, near cloud tops there is a decrease in the droplet concentrations. (iii) TRK5 had the lowest LWC value (below 1.5 $g\,kg^{-1}$).

As can be seen in every cloud σ and $\langle r \rangle$ changes in a similar manner as a function of height (see the green and the yellow lines in the figure) apart from a clear deviation

ACPD
14, 11153–11176, 2014

at the top of TRK3. This observation suggests that the relative dispersion values ($\varepsilon = \sigma/\langle r \rangle$) are constrained within the cloud. This will be examined further.

Figure 3 shows the relation of the relative dispersion with height (Fig. 3a) and with LWC (Fig. 3b). Figure 3a presents the relative dispersion as a function of height with the colored data points representing LWC. It shows that the average relative dispersion changes very little as a function of height. Figure 3b shows the relative dispersion as a function of LWC. The colored data points represent the droplet concentrations. The red lines represent the average value of the relative dispersion. For estimating the average relative dispersion (red lines), the data was divided into ten bins based on the LWC values (in Fig. 3a) or height (in Fig. 3b) and for each bin the average relative dispersion was calculated. While it is clearly evident from the colors of the data points, that on the average for each flight the droplet concentration increases with LWC, the average relative dispersion falls within an almost constant range and it does not depend on LWC. Figure 4 is similar to Fig. 3, but it is based only on measurements in the clouds' boundaries. Figure 4 demonstrates that as for the inner parts of the clouds also in the cloud boundaries the droplet concentration increases with LWC, while the average relative dispersion remains almost constant. Moreover, apart from differences in the total number of data points, the results near the cloud boundary (entrainment zones) are similar to those near the inner parts (more adiabatic). This may suggest that inhomogeneous mixing dominated the entrainment processes under the studied conditions.

It can also be noted that the trend for the TRK3 case is different. A clear decrease in ε is observed near the top of the cloud associated with higher LWC values. Such a trend is likely to be an artifact and will be discussed in the paper's summary.

Figure 5 presents ε as a function of $\langle r \rangle$. The presented $\langle r \rangle$ values are binned in a way that each point represents a different height as shown in Fig. 2. Such representation suggests that the relative dispersion is invariant to changes in the average radius (or to the height within the cloud), reinforcing the conclusions shown on Figs. 3 and 4.

ACPD
14, 11153–11176, 2014

As a final step of the analysis we look at the total mean and variance of ε. Figure 6 shows the ε and its variance values as a function of LWC (Fig. 6a) and droplet concentration (Fig. 6b) using all cases except TRK3. The gray crosses represent ε values, while the blue circles and red crosses represent the binned (number based) mean of ε and the standard deviation of ε.

The results show that only in cloud segments with very low LWC and low drop concentrations (LWC $<\sim 0.01\,\mathrm{g\,kg^{-1}}$ and Nc $<\sim 5\,\mathrm{cm^{-3}}$) ε values significantly vary ($\varepsilon \sim 0.1$–1.25). For higher values of LWC and Nc the values of ε fall within a relatively narrow range of values between 0.24–0.37. The convergence of ε with higher LWC and is further investigated in Fig. S1, while its convergence with aerosol concentration and along the vertical depth of the cloud are further investigated by Figs. S2 and S3, respectively (see Supplement). The convergence of ε with relatively higher LWC and Nc is further reinforced in Fig. S1 while Figs. S2 and S3 further indicate that ε is less dispersive for lower aerosol loading and within or near the cloud base. In a similar manner, the ε variance was shown to be smaller for samples with: higher LWC, within or near the cloud base, and for higher aerosol loading, which again may be associated with higher LWC and Nc.

4 Discussion and summary

Using in situ measurements of droplet size distributions of warm cumulus clouds we investigated the properties of the droplet relative dispersion and its relation to the microphysical properties of clouds.

The results suggest that the mean relative dispersion of such cumulus clouds does not show any significant trend with LWC, height within the cloud, Nc, aerosol loading and with $\langle r \rangle$. On the other hand a second order effect on the ε distribution is clearly seen. Specifically, the variance of ε decreases significantly as the LWC and Nc increase.

ACPD
14, 11153–11176, 2014

Overall the mean relative dispersion values vary between 0.24–0.37 in agreement with previous studies which indicated that ε tends to be bounded in similar narrow range when averaged, for instance in warm cumuli (Pandithurai et al., 2012; Berg et al., 2011), in stratus clouds (Peng et al., 2007) and stratocumulus clouds (Pawlowska et al., 2006). The more scattered ε values (ranging ~ 0.1–1.25) were associated with very low LWC and low Nc, below threshold values of $\sim 0.01\,\mathrm{g\,kg^{-1}}$ and Nc $\sim 5\,\mathrm{cm^{-3}}$. In those regions in clouds the measurement quality is low and this may be reflected as an increase in the ε variance. Other reasons may be associated with either the stage of cloud development or the entrainment within the cloud (Tas et al., 2012), which may explain similar scattering in ε values for low LWC and NC values (e.g., by Pandithurai et al., 2012; Zhao et al., 2006; Deng et al., 2009).

Once above the threshold levels of Nc and LWC, ε showed fast convergence into averaged values. Deng et al. (2009) and Zhao et al. (2006) also indicated convergence of ε to a narrow range (0.4–0.5) with increasing Nc, which was associated with higher pollution levels. Tas et al. (2012) showed that for the core of a cumulus cloud in its mature developing stage, and with high LWC, ε falls into a narrow range. In the present study we also observed convergence of ε with aerosol loading which may be related to an increase in Nc and/or LWC. This increase in number concentration could extend the lifetime of the mature stage, as suggested by the second indirect effect (Albrecht et al., 1989).

How reliable are the ε estimations based on the Cloud Droplet Probe measurements? To represent the ε values right, one needs a full description of the droplet size distribution. Our measurements are limited to the radius range of 1.5–25 µm. Clearly ε will not be represented well in cases when the tail of the size distribution exceeds 25 µm in radius. The distribution for each cloud and vertical level is shown in Fig. 2 and it is evident that except for case TRK3, the concentration of droplets > 20 µm is negligible. The relative dispersion results for the case of TRK3 tend to be smaller for higher LWC, higher up in the cloud. As indicated above, TRK3 was the cleanest case and it probably contained larger drops that were not measured by the probe. This suggests that in

ACPD
14, 11153–11176, 2014

such a case the estimation of ε might be incorrect. Specifically, the contribution of the larger droplets is expected to be more significant for the case of larger LWC higher up in the cloud (see case TRK3 in Fig. 2). Therefore the decrease in ε for such data points might be an artifact due to incomplete representation of the drop size distribution that did not include the contribution of larger drops.

Our analysis suggests that a bias in ε due to a failure to detect the entire droplet size distribution including the tail of large droplet may serve as a criterion for the reliability of using measurement data for microphysical analyses. Currently we are trying to validate this hypothesis using data from other experiments.

Based on the relatively small $\langle r \rangle$ values (see Fig. 2), the sparse population of large droplets (for all clouds except TRK3) and the relatively high aerosol loadings we assume that the growth of the drops was dominated by the condensation process. It is well known that growth by condensation leads to an increase in $\langle r \rangle$ but to a decrease of the width of the size distribution (smaller σ) (e.g. Rogers and Yau, 1989). However, the invariant nature of ε values in the present study and in some of the other studies suggests that additional processes occur simultaneously with the condensation. These additional processes act to increase σ such that the ratio of σ to $\langle r \rangle$ remains relatively constant. Such processes may include the growth of drops by collision-coalescence or the formation of new droplets by activation of cloud condensation nuclei (CCN) (increasing the number of the smaller droplets) or giant CCN (which may increase the number of the larger drops). Both scenarios act to broaden the droplet spectra. In this study, we cannot determine which of these processes is more significant. Moreover, the contribution of each of the two processes, to maintain relatively constant range of ε may vary at different locations and stages of cloud evolution. Collection based processes should be more important higher in the cloud and later in the cloud evolution, while activation of new particles should be more important near cloud base and in the early stages of development.

Currently, autoconversion and radiation parameterizations in GCMs are based on the estimated impact of aerosol loading on the magnitude of ε (see Sect. 1). The present

ACPD
14, 11153–11176, 2014

study that analyze airplane measurements indicates that ε variance, rather than the values of ε, depends on microphysical and thermodynamic processes. This should be further tested under different conditions, in order to improve current parameterization in GCMs.

Supplementary material related to this article is available online at http://www.atmos-chem-phys-discuss.net/14/11153/2014/acpd-14-11153-2014-supplement.pdf.

Acknowledgements. This work was supported by the Israel Science Foundation (grant 1172/10). CARI was funded by the Istanbul Metropolitan Municipality (IMM) through Seeding Operations and Atmospheric Research (SOAR).

References

Ackerman, A. S., Toon, O. B., Taylor, J. P., Johnson, D. W., Hobbs, P. V., and Ferek, R. J.: Effects of aerosols on cloud albedo: evaluation of Twomey's parameterization of clouds susceptibility using measurements of ship tracks, J. Atmos. Sci., 57, 2684–2695, 2000.

Altaratz, O., Koren, I., Remer, L. A., and Hirsch, E.: Review: cloud invigoration by aerosols – coupling between microphysics and dynamics, Atmos. Res., 140, 38–60, 2014.

Albrecht, B.: Aerosols, cloud microphysics, and fractional cloudiness, Science, 245, 1227–1230, 1989.

Andreae, M. O., Rosenfeld, D., Artaxo, P., Costa, A. A., Frank, G. P., Longo, K. M., and Silva-Dias, M. A. F.: Smoking rain clouds over the Amazon, Science, 303, 1337–1342, doi:10.1126/science.1092779, 2004.

Axisa, D., Rosenfeld, J., Santarpia, W., Woodley, and Collins, D.: The Southern Plains Experiment in Cloud Seeding of Thunderstorms for Rainfall Augmentation (SPECTRA) project: operational tools used towards verifying glaciogenic and hygroscopic seeding conceptual models, case studies and preliminary results, Preprints, 16th Conf. on Planned and Inadver-

tent Weather Modification, 9–13 January 2001, San Diego, CA, Amer. Meteor. Soc., Paper 6.6, 2005.
Berg, L. K., Berkowitz, C. M., Barnard, J. C., Senum, G., and Springston, S. R.: Observations of the first aerosol indirect effect in shallow cumuli, Geophys. Res. Lett., 38, L03809, doi:10.1029/2010GL046047, 2011.
Daum, P. and Liu, Y.: Dispersion of cloud droplet size distributions, cloud parameterizations, and indirect aerosol effects, paper presented at 13th ARM Science Team Meeting, Atmos. Radiat. Meas. Clim. Res. Facility, US Dep. of Energy, Daytona Beach, Fla., 14–18 March 2003.
Deng, Z., Zhao, C., Zhang, Q., Haung, M., and Ma, X.: Statistical analysis of properties and parameterization of effective radius of warm cloud in Beijing area, Atmos. Res., 93, 888–896, 2009.
Feingold, G., Eberhard, W. L., Veron, D. E., and Previdi, M.: First measurements of the Twomey indirect effect using ground-based remote sensors, Geophys. Res. Lett., 30, 1287, doi:10.1029/2002GL016633, 2003.
Hansen, J. E. and Travis, L. D.: Light-scattering in planetary atmospheres, Space Sci. Rev., 16, 527–610, doi:10.1007/BF00168069, 1974.
Hsieh, W. C., Nenes, A., Flagan, R. C., Seinfeld, J. H., Buzorius, G., and Jonsson, H.: Parameterization of cloud droplet size distributions: comparison with parcel models and observations, J. Geophys. Res., 114, D11205, doi:10.1029/2008JD011387, 2009.
Koren, I., Kaufman, Y. J., Rosenfeld, D., Remer, L. A., and Rudich, Y.: Aerosol invigoration and restructuring of Atlantic convective clouds, Geophys. Res. Lett., 32, L14828, doi:10.1029/2005GL023187, 2005.
Liu, Y. and Daum, P. H.: Spectral dispersion of cloud droplet size distributions and the parameterization of cloud droplet effective radius, Geophys. Res. Lett., 27, 1903–1906, doi:10.1029/1999GL011011, 2000a.
Liu, Y. and Daum, P. H.: Which size distribution function to use in parameterization of effective radius, paper presented at 13th International Conference on Clouds and Precipitation, 14–18 August 2000, Int. Comm. on Clouds and Precip. Reno, Nev., 2000b.
Liu, Y. and Daum, P. H.: Anthropogenic aerosols: indirect warming effect from dispersion forcing, Nature, 419, 580–581, doi:10.1038/419580a, 2002.
Liu, Y., Daum, P. H., and McGraw, R.: Size truncation effect, threshold behavior, and a new type of autoconversion parameterization, Geophys. Res. Lett., 32, L11811, doi:10.1029/2005GL022636, 2005.

ACPD
14, 11153–11176, 2014

The relative dispersion of cloud droplets

E. Tas et al.

Liu, Y., Daum, P. H., McGraw, R., and Miller, M.: Generalized threshold function accounting for effect of relative dispersion on threshold behavior of autoconversion process, Geophys. Res. Lett., 33, L11804, doi:10.1029/2005GL0255000, 2006a.

Liu, Y., Daum, P. H., and Yum, S. S.: Analytical expression for the relative dispersion of the cloud droplet size distribution, Geophys. Res. Lett., 33, L02810, doi:10.1029/2005GL024052, 2006b.

Liu, Y., Daum, P. H., Guo, H., and Peng, Y.: Dispersion bias, dispersion effect and aerosol-cloud conundrum, Environ. Res. Lett., 3, 045021, doi:10.1088/1748-9326/3/4/045021, 2008.

Lu, M.-L. and Seinfeld, J. H.: Effect of aerosol number concentration on cloud droplet dispersion: a large-eddy simulation study and implications for aerosol indirect forcing, J. Geophys. Res., 111, D02207, doi:10.1029/2005JD006419, 2006.

Lu, M.-L., Feingold, G., Jonsson, H. H., Chuang, P. Y., Gates, H., Flagan, R. C., and Seinfeld, J. H.: Aerosol-cloud relationships in continental shallow clouds, J. Geophys. Res., 113, D15201, doi:10.1029/2007JD009354, 2008.

Ma, J., Chen, A., Wang, W., Yan, P., Liu, H., Yang, S., Hu, Z., and Lelieveld, J.: Strong air pollution causes widespread haze-clouds over China, J. Geophys. Res., 115, D18204, doi:10.1029/2009JD013065, 2010.

Martin, G. M., Johnson, D. W., and Spice, A.: The measurement and parameterization of effective radius of droplets in warm stratocumulus clouds, J. Atmos. Sci., 51, 1823–1842, doi:10.1175/1520-0469(1994)051<1823:TMAPOE>2.0.CO;2, 1994.

Martins, J. A. and Silva Dias, M. A. F.: The impact of smoke from forest fires on the spectral dispersion of cloud droplet size distributions in the Amazonian region, Environ. Res. Lett., 4, 015002, doi:10.1088/1748-9326/4/1/015002, 2009.

McFarquhar, G. M. and Heymsfield, A. J.: Parameterization of INDOEX microphysical measurements and calculations of cloud susceptibility: applications for climate studies, J. Geophys. Res., 106, 28675–28698, doi:10.1029/2000JD900777, 2001.

Miles, N. L., Verlinde, J., and Clothiaux, E. E.: Cloud droplet size distributions in low-level stratiform clouds, J. Atmos. Sci., 57, 295–311, 2000.

Pandithurai, G., Dipu, S., Prabha, T. V., Maheskumar, R. S., Kulkarni, J. R., and Goswami, B. N.: Aerosol effect on droplet spectral dispersion in warm continental cumuli, J. Geophys. Res., 117, D16202, doi:10.1029/2011JD016532, 2012.

Pawlowska, H., Grabowski, W. W., and Brenguir, J.-L.: Observations of the width of cloud droplet spectra in stratocumulus, Geophys. Res. Lett., 33, L29810, doi:10.1029/2006GL026841, 2006.
Peng, Y. and Lohmann, U.: Sensitivity study of the spectral dispersion of the cloud droplet size distribution on the indirect aerosol effect, Geophys. Res. Lett., 30, 1507, doi:10.1029/2003GL017192, 2003.
Peng, Y., Lohmann, U., Leaitch, R., and Kulmala, M.: An investigation into the aerosol dispersion effect through the activation process in marine stratus clouds, J. Geophys. Res., 112, D11117, doi:10.1029/2006JD007401, 2007.
Rogers, R. R. and Yau, M. K.: A Short Course in Cloud Physics, 3rd edn., Pergamon Press, Oxford, 1989.
Rosenfeld, D. and Feingold, G.: Explanation of the discrepancies among satellite observations of the aerosol indirect effects, Geophys. Res. Lett., 30, 1776, doi:10.1029/2003GL017684, 2003.
Rotstayn, L. D. and Liu, Y.: Sensitivity of the first indirect aerosol effect to an increase of cloud droplet spectral dispersion with droplet number concentration, J. Climate, 16, 3476–3481, doi:10.1175/1520-0442, 2003.
Slingo, A.: A GCM parameterization for the shortwave Radiative properties of water clouds, J. Atmos. Sci., 46, 1419–1427, doi:10.1175/1520-0469, 1989.
Tao, W.-K., Chen, J.-P., Li, Z., Wang, C., and Zhang, C.: Impact of aerosols on convective clouds and precipitation, Rev. Geophys., 50, RG2001, doi:10.1029/2011RG000369, 2012.
Tas, E., Koren, I., and Altaratz, O.: On the sensitivity of droplet size relative dispersion to warm cumulus cloud evolution, Geophys. Res. Lett., 39, L13807, doi:10.1029/2012GL052157, 2012.
Teller, A., Axisa, D., Bruintjes, R., Collins, D., Mutlu, A., Kluzek, C., Pocernich, M., Buseck, P. R., Freney, E., Tessendorf, S., Hunter, S., Şen, O., Kocak, K., Toros, H., Köse, A., and Tunc, M.: Cloud and Aerosol Research in Istanbul Final Report 2007–2008, NCAR/RAL, Boulder, CO, 2008.
Twomey, S.: The influence of pollution on the short wave albedo of clouds, J. Atmos. Sci., 34, 1149–1152, 1977.
Xie, X., Liu, X., Peng, Y., Wan, Y., Yue, Z., and Li., X.: Numerical simulation of clouds and precipitation depending on different relationships between aerosol and cloud droplet spectral dispersion, Tellus B, 65, 1–17, 2013.

Yum, S. S. and Hudson, J. G.: Adiabatic predictions and observations of cloud droplet spectral broadness, Atmos. Res., 73, 203–223, doi:10.1016/j.atmosres.2004.10.006, 2005.
Zhao, C., Tie, X., Brasseur, G., Noone, K. J., Nakajima, T., Noone, K. J., Nakajima, T., Zhang, Q., Zhang, R., Huang, M., Duan, Y., Li, G., and Ishizaka, Y.: Aircraft measurements of cloud droplet spectral dispersion and implications for indirect aerosol radiative forcing, Geophys. Res. Lett., 33, L16809, doi:10.1029/2006GL026653, 2006.

ACPD
14, 11153–11176, 2014

Table 1. Airborne measurements which were used for the present study. The table indicates for each of the 5 airborne measurements used for the present study the flight date, aerosol loading at the cloud base (see Sect. 3.2) and the abbreviation used for the flight in this paper.

Flight date/Flight time (LT)	Aerosol loading (cm^{-3}) (0.11–3 µm)	Abbreviation
6 Jun 2008/12:00–12:36	1800	TRK1
6 Jun 2008/12:54–13:24	900	TRK2
7 Jun 2008/06:24–07:36	700	TRK3
7 Jun 2008/13:18–13:45	800	TRK4
7 Jun 2008/14:09–14:30	1000	TRK5

ACPD
14, 11153–11176, 2014

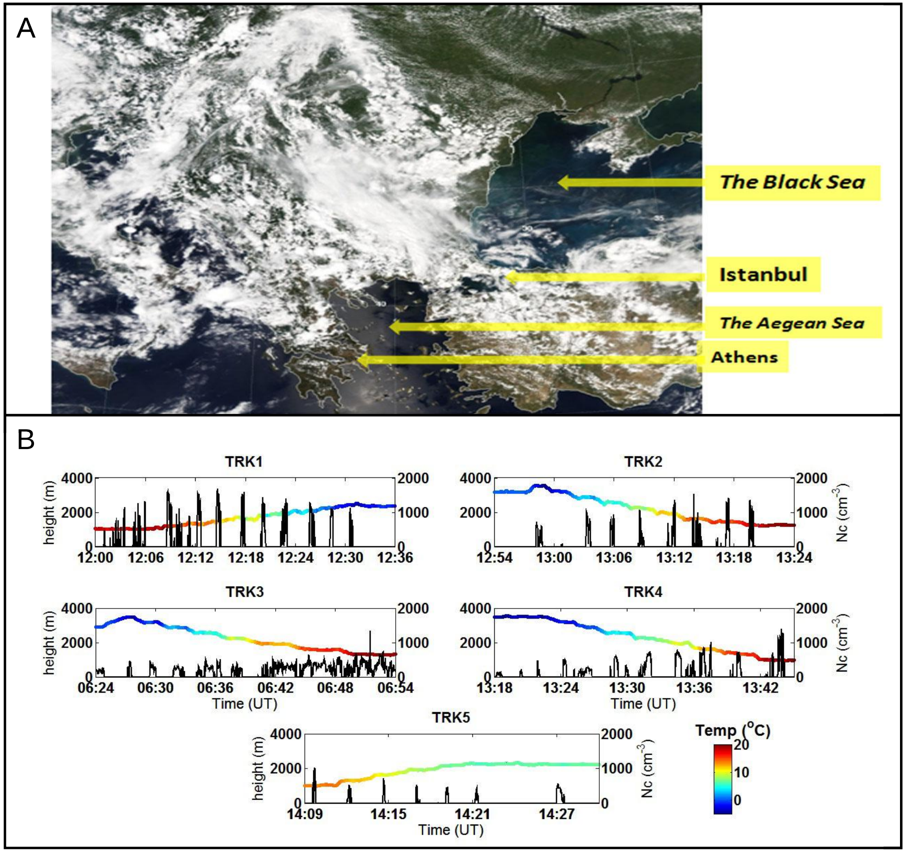

Fig. 1. (A) A MODIS image of the Eastern Mediterranean region on 7 June 2008. **(B)** A summary of flight profiles and cloud droplet concentration of airborne measurements that were carried out in 6–8 June 2008 in Istanbul, Turkey. Black line shows the droplet concentration and the colored line shows the height above ground and the temperature.

ACPD
14, 11153–11176, 2014

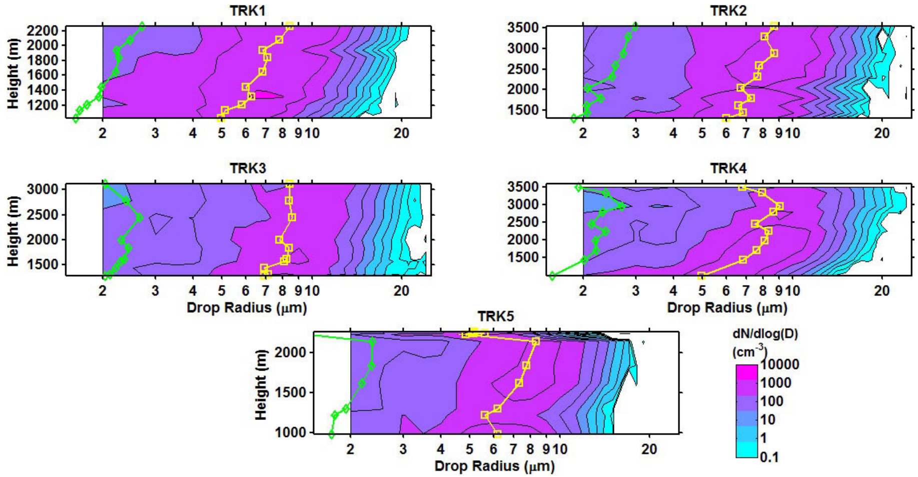

Fig. 2. Cloud droplet size distribution as function of the height above the ground. The colored contours show the concentration ($dN/d\log(D)$) for each radius. The yellow and the green lines represent the average and standard deviation of the radius over the entire measurements, respectively. For the purpose of constructing the lines of the average radius and the standard deviation we divided the measurements into 10 height bins and for each bin the average was calculated.

ACPD
14, 11153–11176, 2014

The relative dispersion of cloud droplets

E. Tas et al.

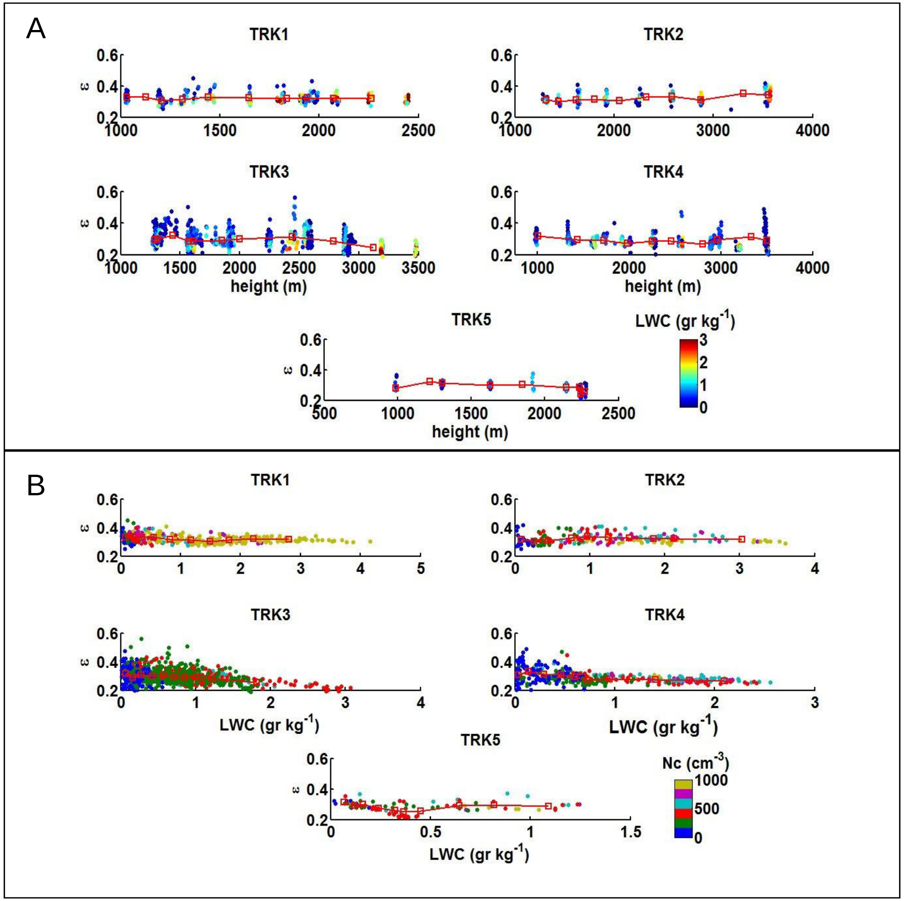

Fig. 3. Relative dispersion (ε) vs. the height above the ground with colors that represent the liquid water content **(A)** and ε vs. the LWC with colors that represent the droplet concentration **(B)** for the in cloud data points.

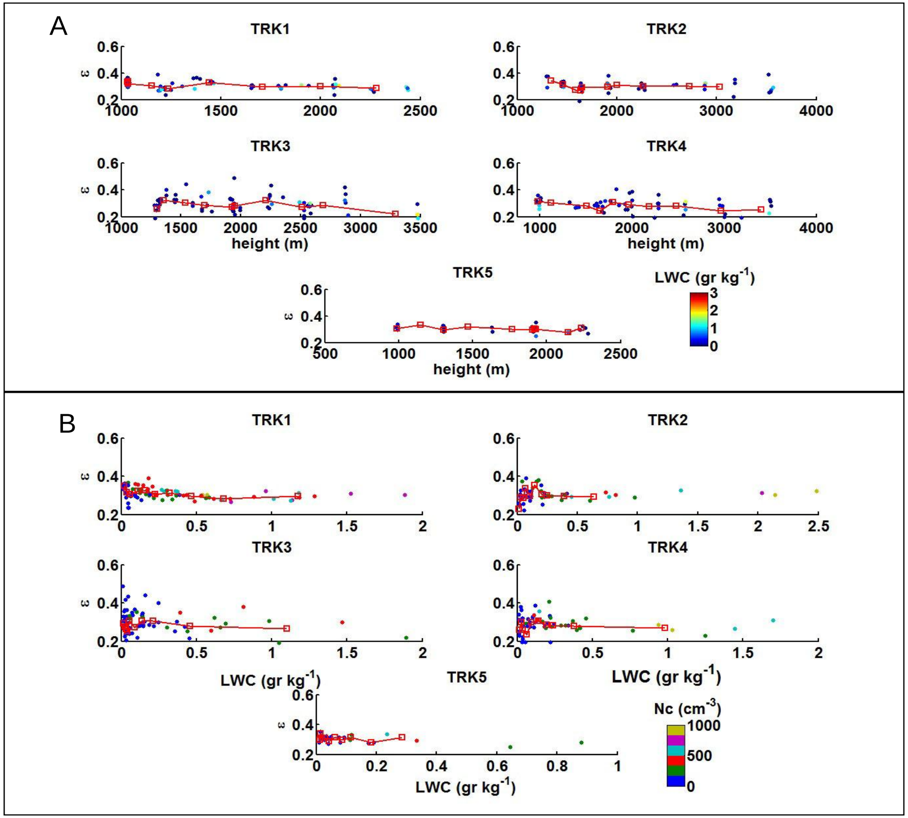

Fig. 4. Relative dispersion (ε) vs. the height above the ground with colors that represent the liquid water content **(A)** and ε vs. the LWC with colors that represent the droplet concentration **(B)** for the cloud boundaries data points.

ACPD
14, 11153–11176, 2014

Title Page
Abstract | Introduction
Conclusions | References
Tables | Figures
◀| |▶
◀ ▶
Back | Close
Full Screen / Esc
Printer-friendly Version
Interactive Discussion

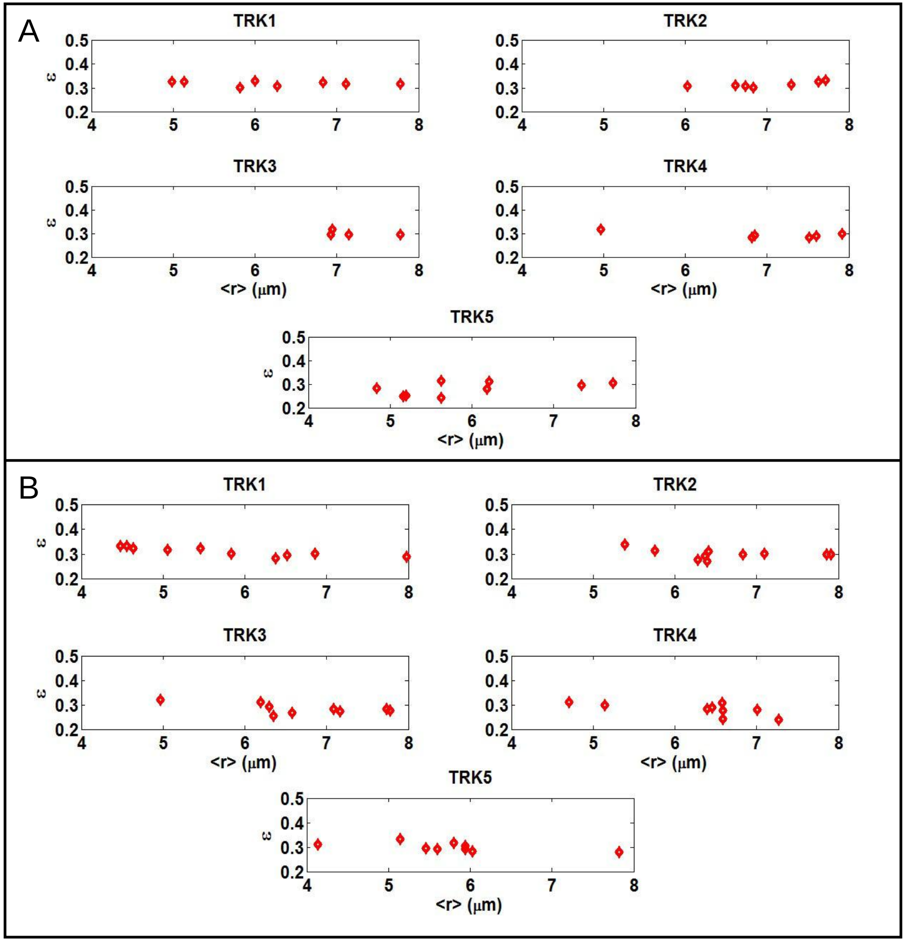

Fig. 5. Relative dispersion vs. average radius. The upper panel **(A)** for the in cloud data and the lower **(B)** for the cloud boundaries.

ACPD
14, 11153–11176, 2014

Title Page
Abstract | Introduction
Conclusions | References
Tables | Figures
◀| |▶
◀ ▶
Back | Close
Full Screen / Esc
Printer-friendly Version
Interactive Discussion

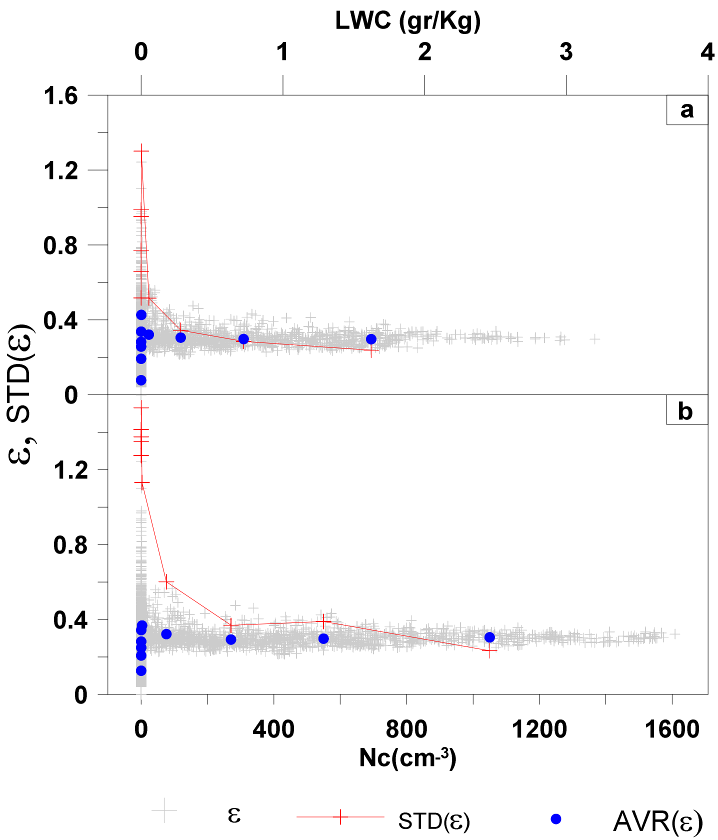

Fig. 6. Relative dispersion and its variance as a function of cloud water liquid content and cloud droplet number. Relative dispersion (ε), relative dispersion average (AVR(ε)) and relative dispersion variance (STD(ε)) are presented vs. LWC **(a)** and Nc **(b)**. AVR(ε) and (STD(ε)) are presented as the average value of number-based 10 size bins.