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Vertical variation of ice particle size in convective cloud tops

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Abstract

A novel technique is used to estimate derivatives of ice effective radius with respect to height near convective cloud tops (dr_e/dz) from airborne shortwave reflectance measurements and lidar. Values of dr_e/dz are about $-6 \mu\text{m}/\text{km}$ for cloud tops below the homogeneous freezing level, increasing to near $0 \mu\text{m}/\text{km}$ above the estimated level of neutral buoyancy. Retrieved dr_e/dz compares well with previously documented remote sensing and in situ estimates. Effective radii decrease with increasing cloud top height, while cloud top extinction increases. This is consistent with weaker size sorting in high, dense cloud tops above the level of neutral buoyancy where fewer large particles are present, and with stronger size sorting in lower cloud tops that are less dense. The results also confirm that cloud-top trends of effective radius can generally be used as surrogates for trends with height within convective cloud tops. These results provide valuable observational targets for model evaluation.

1. Introduction

It is generally observed that ice crystal sizes decrease as a function of altitude within clouds. This dependency is often explained as resulting from size sorting owing to the greater fall speeds of larger particles [Heymsfield, 1975; McFarquhar and Heymsfield, 1997; Yuan and Li, 2010], but may also be related to dependence of ice diffusional growth on available water vapor and temperature [Heymsfield and Donner, 1990] or other factors. Accurate modeling of ice growth and sedimentation is crucial to reliably represent vertical structures of ice clouds and cloud evolution in general [e.g., Sanderson et al., 2007]. Hence, observational targets for the ice size vertical variation are valuable for model evaluation.

Information about the vertical variation of ice sizes can be obtained from in situ measurements, although there are issues with sampling statistics and measurement artifacts [e.g., Lawson et al., 2010; Korolev et al., 2013]. Profiles of ice sizes within clouds can also be estimated from measurements of radar [e.g., Mace, 2002] or combined radar-lidar [e.g., Deng et al., 2010], although such estimates have significant uncertainties because of many unconstrained assumptions [Comstock et al., 2013]. Furthermore, vertical variation of ice

sizes can be estimated from statistics of aircraft or satellite remote sensing retrievals at cloud tops [Yuan et al., 2010; Yuan and Li, 2010; van Diedenhoven et al., 2012a, 2014], although these remote sensing studies use cloud-top trends as surrogates for trends with height within clouds [Lensky and Rosenfeld, 2006, hereafter referred to as time-space interchangeability], which may not always be applicable.

Retrievals of information on ice size vertical variation using effective radii inferred from different shortwave reflectances such as those measured by the Moderate Resolution Imaging Spectroradiometer (MODIS) have been done [e.g., Wang et al., 2009; Zhang et al., 2010]. However, those studies yield estimates of size variation with respect to optical depth into cloud top, rather than with respect to physical depth. Furthermore, using the 1.64 and 2.13- μm channels of MODIS yields little information about vertical structures because of the similar ice absorption in the two channels, while using a combination including the 3.7- μm MODIS band is challenging owing to the influence of thermal emission.

Here we present results of a technique to retrieve trends of effective radius with respect to physical depth within convective cloud tops that uses a combinations of shortwave infrared measurements at 1.59 and 2.26 μm and lidar measurements at 0.532 μm . Ice absorption at 2.26 μm is substantially weaker than at 1.59 μm [Warren and Brandt, 2008], which implies that 2.26- μm measurements generally probe to a deeper optical depth within cloud [Miller et al., 2014]. To estimate the relation between optical depth and physical depth within cloud tops, lidar retrievals of extinction are used. This approach is applied to combined measurements of the Research Scanning Polarimeter [RSP, Cairns et al., 1999, 2003] and the Cloud Physics Lidar [CPL, McGill et al., 2002] mounted on NASA's high-altitude ER-2 aircraft during the Studies of Emissions and Atmospheric Composition, Clouds and Climate Coupling by Regional Surveys (SEAC⁴RS) campaign.

2. Methodology

Optical thickness and ice effective radius are retrieved using visible and shortwave infrared measurements as described by van Diedenhoven et al.[2014]. For the inference of cloud optical thickness, a look-up-table (LUT) is calculated for cloud optical thicknesses in the range 1–20 with steps of 1 and in the range 22–100 with steps of 2. First, the closest match to the measured total reflectance is found within the LUT and then log-linear interpolation is used to estimate the cloud optical thickness for which the simulated reflectance exactly matches the measured reflectance. Here, RSP nadir reflectances at 0.67 and 0.86 μm are used for retrievals over land and water, respectively. The ice optical properties are obtained by assuming mixtures of distorted hexagonal columns and plates with a scattering asymmetry parameter at 0.86 μm of 0.78. These hexagonal prisms serve as radiative proxies for the more complex ice crystals commonly observed. The single-scattering albedo at a given wavelength is mainly determined by the ice crystal effective radius and its sensitivity to particle shape is minimal [Key et al., 2002; van Diedenhoven et al., 2014b]. To obtain optical properties for a range of effective radii, we follow Baum et al. [2005, 2011] and use a large variety of normalized gamma size distributions based on more than 14,000 different in situ measurements in various cloud types [Heymsfield et al., 2013]. Note, however, that effective radius retrievals are not substantially sensitive to the assumption of particle size

distribution shape [Wyser and Yang, 1998; van Dienenhoven et al., 2014]. More details of the optical properties used here are described by van Dienenhoven et al.[2014]. The LUT is calculated at cosine values of solar zenith angles (μ_0) ranging from 0.4 to 1 with steps of 0.1. The LUT values corresponding to μ_0 closest to the median μ_0 for each ER-2 flight leg is selected.

Subsequently, effective radii r_e , defined as three-fourth of the average ice volume divided by their average projected area, are retrieved from nadir reflectances at 2.26 and 1.59 μm . Similarly to the method described in detail by van Dienenhoven et al.[2014], an LUT is used with simulated reflectances of homogeneous ice cloud layers assuming effective radii that range from 5 to 65 μm in steps of 3 μm and cloud optical thicknesses of 0.1, 0.5, 1, 2, 5, 10, 25, 50, 75 and 100. For any given RSP pixel, the LUT is interpolated to the retrieved cloud optical thickness. Then, an effective radius is retrieved from each measurement in a particular band by first determining the closest match within the LUT to the measured nadir reflectance in that band, after which linear interpolation is used to match the simulated and measured reflectances. The same ice optical model is used as for the cloud optical thickness retrievals described above. All measured reflectances are corrected for gaseous absorption [cf. Wu et al., 2015].

To estimate to which optical depth level from cloud top the retrieved effective radii pertain, we use weighting functions as described by Platnick[2000] and Zhang et al.[2010]. For nadir observations, a normalized weighting function w at optical depth level τ within cloud top and at wavelength λ can be defined as

$$w(\lambda, \tau, \mu_0, r_e) = \frac{1}{R(\tau_c, \mu_0, r_e)} \frac{Rd(\lambda, \tau, \mu_0, r_e)}{d\tau}, \quad (1)$$

where τ_c is the total optical thickness of the layer. Here, we calculate w from simulated measurements contained in the LUT described above. The derivative of reflectance with respect to optical thickness is calculated using finite differences at the optical depth values within the LUT. Subsequently, a spline function is fitted to obtain the derivatives at other optical depths, and this interpolated weighting function is renormalized. Figure 1 shows examples of weighting functions at 1.59 and 2.26 μm for an effective radius of 29 μm and at a solar zenith angle of 53° ($\mu_0 = 0.6$). For these calculations $\tau_c=100$, ensuring reflectances are saturated and thus do not increase as optical thickness increases further. The weighting functions for both bands peak at an optical depth between 0.5 and 1, but it is apparent that the 2.26- μm band is more sensitive to deeper layers than the 1.59- μm band [cf. Miller et al., 2014].

The weighting functions can be used to estimate the characteristic optical depth to which a given effective radius retrieval pertains τ_w by

$$\tau_w(\lambda, \mu_0, r_e) = \int_0^{\tau_c} w(\lambda, \mu_0, \tau, r_e) \tau d\tau. \quad (2)$$

For the 1.59 and 2.26- μm bands, τ_w is indicated in Fig. 1. As expected, τ_w for the 2.26- μm band is greater than at 1.59 μm . The difference τ_w between the τ_w for the 1.59 and 2.26- μm bands depends on μ_0 and r_e and ranges from about 0.5 to 1.1. τ_w can be interpreted as a characteristic probing depth difference between the two channels. Thus, defining r_e as the difference in retrieved r_e from the two channels, the derivative of effective radius in cloud top with respect to optical depth can be approximated by $r_e = \tau_w$.

To estimate the derivative of r_e with respect to physical depth in the cloud top z , an estimate of $d\tau/dz$, that is cloud top extinction, is needed. Here, the cloud top extinction σ_{ct} is derived from CPL lidar measurements. CPL estimates the optical depth for each 30-m range bin by assuming or inferring an extinction to backscatter (lidar) ratio [Yorks et al., 2011]. Usually, CPL measurements are fully attenuated at an optical depth of around 3, but the physical penetration depth of the lidar signal depends on the extinction near cloud top. Here, we estimate cloud top extinction by determining the difference in altitude at which certain lower and upper optical depth thresholds, τ_{top} and τ_{max} , respectively, are reached, i.e.,

$$\sigma_{ct} = \frac{d\tau}{dz} = - \frac{\tau_{top} - \tau_{max}}{z(\tau_{top}) - z(\tau_{max})}. \quad (3)$$

Here, we use $\tau_{max} = 2.6$ since it is sufficiently close to the maximum attenuation optical depth of CPL while still most measurements reach this level. We use $\tau_{top} = 0.1$, which avoids influence of optically thin, separate cirrus layers above the convective clouds present on some days of SEAC⁴RS. Finally, for each combined RSP and CPL measurement, the derivative of r_e with respect to physical depth in the cloud top can be estimated by

$$\frac{dr_e}{dz} = \frac{\Delta r_e}{\Delta \tau_w} \sigma_{ct}. \quad (4)$$

Retrievals of r_e are known to be sensitive to the assumed ice model [Yang et al., 2008; Baum et al., 2011]. In particular, for thick clouds retrieved r_e scales with the asymmetry parameter g of the phase function approximately as $(1-g)$ [van Dienenhoven et al., 2014]. Thus, it is expected that r_e also scales as $(1-g)$. Furthermore, the estimate of τ using w is found to scale roughly as $1/(1-g)$, leading to a $(1-g)^2$ dependency of dr_e/dz . Note that RSP is capable of retrieving asymmetry parameter for each pixel [van Dienenhoven et al., 2012b, 2013], but to focus this study on retrieving vertical variations of effective radii, a single ice model with $g = 0.78$ is used for all data. In addition, any variation of ice shape and thus asymmetry parameter with height near cloud top is neglected. Not shown here, but retrieved g values for the SEAC⁴RS campaign range from about 0.76 to 0.78 at cloud top heights around 14 km, implying that using $g = 0.78$ may lead to negative biases in estimated dr_e/dz of 0–20%, while retrieved g range from about 0.78 to 0.82 at around 8 km, indicating possible positive biases in retrieved dr_e/dz of 0–50%. Finally, σ_{ct} and thus dr_e/dz are linearly affected by uncertainties in the lidar ratios, which are determined as described by Yorks et al. [2011] and have an estimated variability of 30–40%. Lidar ratio may depend on particle size, shape and roughness, but these relationships are unconstrained as current calculation

methods are generally unreliable to simulate ice optical properties at exact backscattering [Zhou and Yang, 2015]. No general, systematic relation between estimated lidar ratio and cloud mid-layer temperature or depolarization ratio (a proxy for ice particle shape) was found by Yorks et al. [2011].

3. Data

RSP and CPL data obtained during SEAC⁴RS are used here. SEAC⁴RS was based in Houston in 2013 and targeted the continental United States and the Gulf of Mexico. ER-2 aircraft flight legs over deep convective clouds and fresh outflow from August 21 and 30 and September 2, 4, 11, 13 and 22 are selected from visual inspection of CPL profiles and GOES imagery. Days with substantial influence of apparent advected outflow of mesoscale systems over land are avoided. Interference of liquid clouds is avoided by only selecting RSP pixels for which liquid index values, computed from the RSP multi-directional polarization measurements as described by van Dienenhoven et al. [2012a], are below 0.3. Further, only pixels with an optical thickness exceeding 5 are included in the analysis to avoid influence of the surface and to make sure that $\tau_{\max}$ (Eq. 3) can be determined. In total 11465 pixels are selected. Cloud top heights are determined from RSP's multi-directional reflectances using a multi-angular technique [cf. Alexandrov et al., 2012]. Generally, RSP cloud top heights compare well with CPL determined cloud top heights, as will be demonstrated in future work. The CPL and RSP instruments have similar fields of view and here CPL and RSP measurements with the closest time stamps are combined. RSP field of view is about 14 mrad, leading to pixel sizes of about 50–100 m for aircraft-cloud separations between 4 and 8 km.

Using average sonde profiles from the Southeast America Consortium for Intensive Ozone-sonde Network Study (SEACIONS) network near Houston during the SEAC⁴RS campaign, the homogeneous freezing level (HFL, assumed to be at -38°C here), level of neutral buoyancy [LNB, cf. Folkins, 2002] and the cold point tropopause are estimated to be at about 10.5, 12.5 and 16.5 km, respectively. Aside we note that the HFL commonly coincides roughly with the level of minimum stability in regions of deep convection [Pan et al., 2014] estimated from the sonde data.

4. Results

The retrievals of effective radii are binned in 0.5-km wide cloud top height bins and the median and upper and lower middle quartiles in each bin are shown in Fig. 2. The minimum and maximum number of samples in the cloud top height bins is 117 at 8–8.5 km and 2425 at 12.5–13 km, respectively. Effective radii generally decrease with increasing cloud top height, consistent with previous remote sensing studies [Yuan and Li, 2010; van Dienenhoven et al., 2012a, 2014]. Furthermore, the interquartile ranges of effective radii decrease with increasing cloud top height [cf. van Dienenhoven et al., 2014]. As expected, generally larger effective radii are retrieved using the 2.26- μm band, since it probes deeper than the 1.59- μm band and effective radii are expected to decrease with altitude. In turn, cloud top extinction generally increases with cloud top height as shown in Fig. 3. This implies that at least part of the decrease of effective radius with increasing height is caused

by increasing total particle area rather than decreasing total particle volume (or mass). Maximum cloud top extinction values are obtained just above the estimated LNB, and they decrease again at higher cloud top heights. Since the difference between τ_{top} and τ_{max} here is 2.5, the estimated physical depth over which the lidar data are used is $2.5/\sigma_{\text{ct}}$, and typical values are about 0.8 km at high cloud tops to 2.5 km at lower cloud top heights. Combining these retrievals as explained in section 2 yields estimates of dr_e/dz shown in Fig. 4. Median values of dr_e/dz are about $-6 \mu\text{m}/\text{km}$ below the HFL at 10.5 km, above which it increases until above the estimated LNB at 12.5 km where dr_e/dz values are around $0 \mu\text{m}/\text{km}$. Interquartile ranges of dr_e/dz are about $7\text{--}13 \mu\text{m}/\text{km}$ below 12.5 km and are narrower at around $4\text{--}6 \mu\text{m}/\text{km}$ for higher clouds.

A speculative interpretation of the results is that the high clouds near or above the LNB with dense tops, implying strong extinction, are at or near convective cores where large crystals are lacking because convection may not be able to lift such larger particles to those altitudes and/or particle growth is less efficient at those levels, leading to weak size sorting and consequently small values of dr_e/dz . Note that none of these clouds reach the cold point tropopause level at about 16.5 km, implying substantial entrainment. Clouds at lower altitudes and generally weaker extinction can be interpreted as growing or decaying ("transient") convection or "terminal" cumulus congestus [Luo et al., 2009]. The more negative values of dr_e/dz and larger r_e at such lower altitudes suggest stronger size sorting because of the presence of more larger crystals. Furthermore, crystal aggregation at these levels may increase effective radii as total particle projected area is decreased while total ice volume is conserved. We aim to investigate these possible mechanisms in future work including further analysis of observations as well as simulations. Note that cloud-resolving model simulations evaluated by van Diedenhoven et al. [2012a] produced trends of cloud top effective radii that are similar to observations.

The derivative of effective radius with respect to cloud top height can also be estimated from the median retrieved effective radii within the various cloud top height bins by assuming time-space interchangeability as discussed by Lensky and Rosenfeld[2006] is applicable. Estimates using this approach are also shown in Fig 4 and generally compare favorably with the median values of the retrieved dr_e/dz using the combined CPL and RSP data, especially for the $2.26\text{--}\mu\text{m}$ band, although the estimates disagree for the lowest and highest cloud tops. This is a strong indication that the time-space interchangeability is generally applicable for convective clouds as sampled during SEAC⁴RS within roughly 1–3 km of cloud top, at least for cloud heights between about 10 and 14 km.

Using the time-space interchangeability, dr_e/dz can also be estimated from previous remote sensing studies on the vertical variation of effective radius with respect to cloud height, pressure or temperature. For example, from global MODIS retrievals analyzed by Yuan and Li[2010], we estimate values of dr_e/dz of around -2 to $-3 \mu\text{m}/\text{km}$ at 10–13 km, which fall within the range of median values (-6 to $-1 \mu\text{m}/\text{km}$) of our retrievals at those altitudes. Furthermore, MODIS data in the tropical west Pacific shown by van Diedenhoven et al. [2012a] indicate $dr_e/dz \approx -6 \mu\text{m}/\text{km}$ below the HFL, whereas median values of our retrievals are also about $-6 \mu\text{m}/\text{km}$ there. For high cloud tops, data of van Diedenhoven et al. [2012a] indicate about $-0.5 \mu\text{m}/\text{km}$, which is consistent with our values near $0 \mu\text{m}/\text{km}$ for high cloud

tops. Using a dataset extended from that of van Diedenhoven et al. [2012a], the study by van Diedenhoven et al. [2014] suggests a typical dr_e/dz of about $-1.5 \mu\text{m}/\text{km}$ over a cloud top height range of 9 to 16 km, which is somewhat higher than the median of values above 9 km retrieved in this work, which is $-2.3 \mu\text{m}/\text{km}$. However, one time-period analyzed by van Diedenhoven et al. [2014] showed a substantially stronger trend of effective radius with $dr_e/dz \approx -3 \mu\text{m}/\text{km}$, consistent with the substantial variation seen in our retrievals. Estimates of dr_e/dz can also be derived from previous in situ observations of effective radii at different aircraft altitudes in deep convective clouds. For example, the profiles shown by McFarquhar and Heymsfield [1998] indicate values of dr_e/dz ranging from about -1 to $-4 \mu\text{m}/\text{km}$ at altitudes between 8 and 15 km, while estimates from data by Lawson et al. [2010] fluctuate from -8 to $0 \mu\text{m}/\text{km}$ between 9 and 14 km altitude. These are largely consistent with the range of median trends shown in Fig. 4, which are -6 to $0 \mu\text{m}/\text{km}$ for a similar altitude range. Interestingly, dr_e/dz estimates from data of Lawson et al. [2010] above 14 km exceed $-20 \mu\text{m}/\text{km}$, which is well outside the range of trends retrieved here at high altitudes. Measurements presented by Garrett et al. [2005] suggest dr_e/dz varies substantially from about -6 to $+2 \mu\text{m}/\text{km}$ below about 11 km, while positive values of dr_e/dz are not retrieved from our data at these altitudes. At cloud tops above 13 km the data of Garrett et al. [2005] suggest dr_e/dz values of around $0 \mu\text{m}/\text{km}$ above 13 km, which are consistent with our retrievals. Thus, in general, the estimates from prior remote sensing and in situ measurements compare favorably with our retrievals.

5. Conclusions and Discussion

A novel technique to retrieve the derivative of effective radius with respect to physical depth within cloud tops from combined reflectances at two shortwave infrared wavelengths and lidar measurements is applied to data of the airborne RSP and CPL instruments over deep convective clouds in the southern United States. Values of dr_e/dz are derived from retrieved effective radii, which are found to decrease with increasing cloud top height, and cloud top extinction, which increases with cloud top height up to an altitude of about 12.5 km, corresponding to the estimated level of neutral buoyancy. Values of dr_e/dz are about $-6 \mu\text{m}/\text{km}$ for cloud top heights below 10.5 km, which corresponds to the homogeneous freezing level. Values of dr_e/dz increase with increasing cloud top height between 10.5 km and the level of neutral buoyancy at 12.5 km, above which dr_e/dz values are around $0 \mu\text{m}/\text{km}$. Furthermore, interquartile ranges of dr_e/dz above 12 km are generally narrower than at lower levels. The retrieved dr_e/dz values compare well with those estimated using the statistics of effective radii retrieved from the individual shortwave bands at different cloud tops assuming time-space interchangeability. In general, the estimated values of dr_e/dz compare well with those estimated from previously documented remote sensing and in situ measurements of effective radius at various altitudes. The results are consistent with fewer large particles being present at high cloud tops, leading to weaker size sorting at those altitudes compared to lower cloud top heights.

These results can be used to improve observational constraints on cloud-resolving and climate simulations with prognostic ice number concentrations [e.g., van Diedenhoven et al., 2012a; Kay et al., 2012]. Following the Francis et al. [1994] definition of ice effective radius [cf. McFarquhar and Heymsfield, 1998], such simulations can be readily constrained by

adding to standard model diagnostics (ice number and mass concentration) the ice projected area concentration, which has been suggested as superior to number concentration for comparison of model results with in situ observations owing to the availability of direct measurements with low uncertainty relative to that associated with number or mass [Jackson et al., 2015].

Here, the technique is applied to measurements of CPL and RSP at 1.59 and 2.26 μm . The results can help interpret differences obtained between the ice effective radii retrieved from measurements in similar bands from other platforms, such as Visible/Infrared Imaging Radiometer Suite (VIIRS) on the Suomi National Polar-orbiting Partnership and the future Joint Polar Satellite System satellites (JPSS), as well as the future Himawari-8 Advanced Himawari Imager (AHI) and Geostationary Operational Environmental Satellites (GOES-R) Advanced Baseline Imager (ABI). Furthermore, although the MODIS bands at 1.64 and 2.13 μm do not provide enough contrast to estimate sufficiently accurate derivatives of effective radius with respect to optical depth [Wang et al., 2009; Zhang et al., 2010], a combination of these retrievals with those using the MODIS 3.7- μm band together with measurements of the Cloud-Aerosol Lidar with Orthogonal Polarization (CALIOP) lidar may provide global statistics of derivatives of effective radius with respect to physical depth within cloud tops.

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Key points

- Combinations of shortwave infrared measurements and lidar allow retrievals of trends of effective radius with respect to height in cloud tops.
- Size trends are about -5 micron/km for cloud heights below homogeneous freezing level increasing to 0 micron/km for high cloud tops.
- Results confirm that cloud-top trends of effective radius can generally be used as surrogates for trends with height within convective cloud tops.

Figure 1.

Vertical weighting function examples for RSP's 1.59 (blue) and 2.26- μm (red) channels. Here, r_e is 29 μm and $\mu_0=0.6$. Blue and red dotted lines indicate characteristic probing depths (τ_w) for the 1.59 and 2.26- μm channel, respectively.

Figure 2.

Upper and lower middle quartiles of retrieved effective radii from RSP's 1.59 (blue) and 2.26- μm (red) channels in 0.5-km wide cloud top height bins. The estimated homogeneous freezing level (HFL) and level of neutral buoyancy (LNB) are indicated by dotted lines.

Figure 3.
Upper (dark grey) and lower (light grey) middle quartiles of cloud top extinction.

Figure 4.

Same as Fig. 3, but for dr_e/dz . Solid lines show profiles of dr_e/dz estimated from RSP's 1.59 (blue) and 2.26- μm (red) channels using the time-space interchangeability (TSI) assumption.