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The interdependence of continental warm cloud properties derived from unexploited solar background signals in ground-based lidar measurements

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Abstract. We have extensively analysed the interdependence between cloud optical depth, droplet effective radius, liquid water path (LWP) and geometric thickness for stratiform warm clouds using ground-based observations. In particular, this analysis uses cloud optical depths retrieved from untapped solar background signals that are previously unwanted and need to be removed in most lidar applications. Combining these new optical depth retrievals with radar and microwave observations at the Atmospheric Radiation Measurement (ARM) Climate Research Facility in Oklahoma during 2005–2007, we have found that LWP and geometric thickness increase and follow a power-law relationship with cloud optical depth regardless of the presence of drizzle; LWP and geometric thickness in drizzling clouds can be generally 20–40% and at least 10% higher than those in non-drizzling clouds, respectively. In contrast, droplet effective radius shows a negative correlation with optical depth in drizzling clouds and a positive correlation in non-drizzling clouds, where, for large optical depths, it asymptotes to 10 µm. This asymptotic behaviour in non-drizzling clouds is found in both the droplet effective radius and optical depth, making it possible to use simple thresholds of optical depth, droplet size, or a combination of these two variables for drizzle delineation. This paper demonstrates a new way to enhance ground-based cloud observations and drizzle delineations using existing lidar networks.

1 Introduction

The response of global mean surface temperature to emissions of greenhouse gases from human activities remains highly uncertain (e.g. Hawkins and Sutton, 2009). One of the primary sources of the uncertainty is how low-topped boundary-layer clouds will respond to the temperature perturbation and subsequently amplify or dampen climate change (e.g. Bony and Dufresne, 2005; Bony et al., 2006). To improve representations of cloud properties and their interactions with radiation and water budget in models, sustained efforts have been made to observe and study marine low-topped clouds (e.g. Martin et al., 1994; Kubar et al., 2009; Bretherton et al., 2010; Wood, 2012). However, similar efforts have not been made for mid-latitude continental stratus and stratocumulus clouds, despite their strong links to local weather and climate (Del Genio and Wolf, 2000; Kollias et al., 2007), and their high occurrences compared to other cloud types over land (Sassen and Wang, 2008).

Ground-based observations for mid-latitude continental clouds are primarily provided by the Atmospheric Radiation Measurement (ARM) Climate Research Facility (Stokes and Schwartz, 1994), the NASA Aerosol Robotic Network (AERONET; Holben et al., 1998), the European project Cloudnet (Illingworth et al., 2007) and its descendant AC-TRIS (Aerosols, Clouds, and Trace gases Research Infrastructure Network). At the ARM Oklahoma site, low stratus clouds have been investigated in a variety of studies, from short-period field campaigns along with airborne and/or spaceborne measurements (Sassen et al., 1999; Dong et al.,...
2.1 Retrieving cloud optical depth from calibrated solar background light

Prior to July 2006, the micropulse lidar (MPL) at the ARM Oklahoma site was operated at a wavelength of 523 nm and provided unpolarized measurements at 30 s intervals. Since July 2006, the lidar operated at 532 nm with polarized measurements at 3–10 s temporal resolution. The FOV is 50 µrad. Solar background light is estimated from the averaged signal at lidar range gates between 45 and 55 km, where the molecular backscatter is negligible, and is calibrated against principal plane measurements from AERONET to account for lidar filter degradation and window cleanliness.

Note that, for sites where collocated AERONET measurements are unavailable, one can calibrate solar background light by capitalising on the optical depth of thin clouds retrieved from active lidar signals. Specifically, radiance can be calculated through radiative transfer using thin cloud properties as input, and then be further used to calibrate the corresponding measured solar background light. Details of this alternative calibration approach can be found in Yang et al. (2008).

2.1 Retrieving cloud optical depth from calibrated solar background light

Solar background light received by a lidar is a function of cloud optical depth, cloud effective radius, cloud fraction, surface albedo and solar zenith angle. Figure 1a shows that calibrated solar background light increases with cloud optical depth for optically thin clouds due to increasing scattering of solar radiation into the FOV, and decreases for optically thick clouds due to increasing attenuation, resulting...
cloud optical depth; the final cloud optical depth is reported as the mean and standard deviation of these 40 retrievals. The choice of 40 repetitions is arbitrary, but it affects retrievals insignificantly by 2% compared to results from 1000 repetitions (Chiu et al., 2012). The overall retrieval uncertainty in cloud optical depth is ~10%. Note that, with an uncertainty of 10% rather than 5% in calibrated solar background light, the overall retrieval uncertainty in cloud optical depth will increase to 17–25%.

Since the relationship between zenith radiance and cloud optical depth is not monotonic, the aforementioned retrieval process results in two possible solutions at a given radiance: one corresponds to optically thin clouds, the other corresponds to optically thick. To remove this retrieval ambiguity, Chiu et al. (2007) applied a manual screening. Here we have developed an objective discrimination method using lidar backscatter measurements. We calibrated lidar backscatter signals in clear-air periods using the known molecular scattering at the lidar wavelength. Since the lidar energy was monitored and the lidar optics were assumed to not vary significantly, calibration coefficients from a suitable clear-air period were then extrapolated into cloudy periods. Figure 1b shows an example of the vertical profiles of calibrated attenuated backscatter signals for optically thin and thick clouds. For thick clouds, the attenuated backscatter signal drops dramatically above the apparent cloud top; the mean logarithm (base 10) of the lidar signal from the cloud top to the layer 1 km above is around −7.5. In contrast, for optically thin clouds, the mean logarithm value above cloud tops is around −6.0. The difference between these two mean values is significant, suggesting that this parameter can be used to discriminate between optically thin and thick clouds; however, a proper threshold needs to be determined objectively, as described next. For convenience, the mean of the lidar attenuated backscatter signal from the apparent, or detectable, cloud top to the level 1 km above is denoted as $\beta_{\text{ct},1\text{km}}$.

The threshold of $\beta_{\text{ct},1\text{km}}$ for discriminating cloud optical depth was determined through cases selected objectively using retrievals from shortwave narrowband irradiance measurements (Min and Harrison, 1996a, b), available in the ARM Archive. These cases represent clear or optically thin clouds, selected when the irradiance-based cloud optical depths were smaller than 5 for at least 60 consecutive minutes. The threshold of optical depth 5 was chosen because the zenith radiance typically peaks at this optical depth, and because the lidar signal tends to be completely attenuated beyond this value. For ARM unpolarized lidar measurements, Fig. 2 shows that $\beta_{\text{ct},1\text{km}}$ values range between −8.2 and −5.6, and 94% of cases have values of $\beta_{\text{ct},1\text{km}}$ greater than −7.0. For ARM polarized measurements, the threshold $\beta_{\text{ct},1\text{km}}$ of −6.8 successfully captures a similar fraction (95%) of clear-sky cases. Since this threshold does not vary much over time, we then used $\beta_{\text{ct},1\text{km}}$ thresholds of −7.0 and −6.8 for unpolarized and polarized measurements, respectively, throughout the entire analysis.

Finally, since our lookup tables were based on liquid water clouds, ice clouds were excluded using the lidar depolarization ratio and cloud base height. Based on 5-year ground-based lidar and radiosonde measurements, Naud et al. (2010) suggested a depolarization ratio threshold of 11% for differentiating ice from liquid. We found that this threshold...
generally worked well, but occasionally missed ice clouds when cloud bases were high or clouds were not sufficiently thick. To mitigate these issues, a second criterion involving cloud base height was applied. Based on airborne lidar measurements, Hogan et al. (2004) conducted a global investigation of stratiform supercooled liquid water clouds and showed that less than 10% of supercooled liquid water clouds occurred at temperatures colder than $-20^\circ C$. This temperature threshold approximately corresponds to an altitude of 7 km at the ARM Oklahoma site during summer seasons; any clouds located higher than 7 km were excluded and not retrieved in this study. When lidar depolarization ratio was not available, we used merged sounding data and excluded cases with apparent cloud tops (identified by lidar) above the freezing level. Note that these exclusion criteria are simple yet imperfect, particularly when clouds are thick and lidar cannot detect the true cloud top. Therefore, we further excluded time periods when 1 min ice water paths (IWP) were greater than zero, based on retrievals from the Cloudnet algorithm that uses empirical relationships between ice water content, radar reflectivity and temperature (Hogan et al., 2006).

### 2.2 Calculating cloud effective radius and discriminating drizzling clouds

Once cloud optical depth is retrieved, cloud effective radius can be estimated by combining liquid water paths (LWPs) with two commonly used approaches. The first assumes a constant effective radius in the vertical (Stephens, 1978), and the second assumes a constant cloud droplet number concentration and a linear increase of liquid water content in the vertical (Wood and Hartmann, 2006). Using simultaneous retrievals of cloud optical depth and effective radius at the ARM Oklahoma site, Chiu et al. (2012) found that the second assumption led to a better agreement with LWP measured by microwave radiometers (MWRs) in all sky conditions. Thus, we estimated cloud effective radius $r_{\text{eff}}$ by

$$ r_{\text{eff}} = \frac{9}{5} \frac{\text{LWP}}{\rho_w \tau}, $$

where $\rho_w$ is the density of water, and $\tau$ is cloud optical depth. LWP retrievals are available in the ARM Archive MWRRET product with an uncertainty of 20–30 g m$^{-2}$ and a 20 s time resolution, based on Turner et al. (2007) using two-channel microwave radiometers.

To investigate how the interdependence of cloud macrophysical and microphysical properties on $\tau$ differs between non-drizzling and drizzling clouds, we used the ARM Active Remotely Sensed Clouds Locations (ARSCCL; Clothiaux et al., 2000) product for estimating cloud geometric thickness and for diagnosing drizzling clouds. Combining measurements of cloud radar, micropulse lidar, and ceilometer, ARSCL provides cloud boundary heights and reflectivity at 10 s resolution and 45 m vertical resolution. Cloud geometric thickness was derived from the lowest cloud base (typically detected by lidar) and the cloud top height (detected by radar). We restrict our analysis to single-layer warm clouds by selecting cases with geometrical thicknesses less than 1.5 km, minimising cases of multilayer precipitating clouds that are hard to separate by radar reflectivity and could be erroneously identified as a single-layer cloud. When clouds were sufficiently thick and no significant radar returns were detected, no valid geometric thickness could be obtained and thus such clouds were omitted in our analysis.

Additionally, drizzle discrimination was based on radar reflectivity ($Z$) at the lowest cloud base. Similar to Suzuki et al. (2011), we identify clouds as “non-drizzling” if $Z$ is less than $-15$ dBZ, and “drizzling” if $Z$ is greater than $-15$ dBZ. According to the relationship $R = 0.0788 \times Z^{0.75}$ (rain rate $R$ in mm h$^{-1}$ and $Z$ in mm$^2$ m$^{-3}$) derived from data in Mann et al. (2014), this threshold of $-15$ dBZ corresponds to $\sim 0.006$ mm h$^{-1}$.

### 3 Evaluation of optical depth retrievals

We evaluate our retrievals against a number of benchmarks. The first benchmark is retrievals using zenith radiances from AERONET cloud-mode observations that provide unambiguous cloud optical depth by capitalising on the surface reflectance contrast between 440 and 870 nm wavelengths (Chiu et al., 2010). This benchmark works for all-sky conditions, but retrievals are available only when clouds block the Sun, so AERONET sun photometers operate in cloud mode rather than normal aerosol mode. Cloud-mode retrievals (level 1.5) are available on the AERONET web page.

The second benchmark is retrievals from irradiance measurements at 20 s temporal resolution, available in the ARM Archive. This method uses direct and diffuse transmittance at 415 nm to estimate cloud optical depth with a default
Figure 3. (a) Attenuated backscatter signals from micropulse lidar on 19 April 2005. (b) Time series of cloud optical depth retrieved from a microwave radiometer (MWR), lidar solar background signals from micropulse lidar (MPL), AERONET cloud-mode observations, and ARM Min product (using narrowband irradiance measurements). MWR-based retrievals (grey lines) are based on an assumed cloud effective radius of 8 µm; grey error bars denote lower and upper limits for MWR values, corresponding to a change in droplet effective radius from 6 to 14 µm.

The third benchmark is retrievals using LWP in the ARM Archive MWRRET product (see Sect. 2.2) and Eq. (1) with an assumed effective radius of 8 µm, a typical value for the Oklahoma site (Kim et al., 2003). However, since the true cloud effective radius is not necessarily 8 µm, we further estimate the potential range of cloud optical depth by varying effective radius from 6 to 14 µm. Clearly, retrieval comparison to the third benchmark is intended to qualitatively evaluate cloud optical depth variations, rather than a quantitative measure.

In Sect. 3.1, we present intercomparison results from case studies, including broken cloud and overcast cloud scenes. Additionally, since irradiance-based retrievals work best for homogenous scenes, we focus on overcast stratiform clouds during the period 2005–2007 in Sect. 3.2.

3.1 Case study

Figure 3 shows time series of lidar backscatter signals and cloud optical depths on 19 April 2005 at the ARM Oklahoma site. The penetrated signal at 17:00 UTC and the completely attenuated signal at 20:30 UTC indicate the presence of clear-sky and thick clouds, respectively. These indications of cloud presence by active lidar signals in Fig. 3a correspond well to optical depth retrievals in Fig. 3b. Figure 3b also shows that retrievals from calibrated lidar solar background light agree with those from AERONET cloud mode and from microwave observations for intermittent and broken cloud situations between 17:00 and 18:00 UTC.

Examining two more cases on 10 April and 2 May 2007 when both non-drizzling and drizzling periods are apparent, Figs. 4 and 5 show consistent agreements between our retrievals and the benchmark retrievals. Note that ∼20% of clouds between 14:00 and 18:00 UTC in Fig. 5 are multilayered. Since drizzle classification is based on cloud-base reflectivity, these multilayer clouds are excluded in the following analyses to ensure that LWP, geometric thickness, optical depth and drizzling characteristics refer to the same lowest cloud layer. In short, the overall agreement between independent retrievals suggests that the calibration of solar background light and the newly developed method for distinguishing thin and thick clouds work well for all-sky conditions.

3.2 Stratiform clouds during the period 2005–2007

This section reports results of intercomparison between retrievals from lidar solar background, AERONET cloud-mode observations, and from narrowband irradiance measurements for relatively homogenous and overcast cloud
cases. To objectively select appropriate low-level stratiform water clouds, combined measurements from cloud radar, micropulse lidar, and ceilometer in the ARSCL product were used to identify 1 h time periods with cloud fraction greater than 0.95 and cloud top heights lower than 5 km. Since our analysis includes several data sets at various temporal resolutions, we average data points over a 1 min time period. We took a simple linear average for LWP retrievals and radar reflectivity, but used a logarithm averaging technique for lidar-based cloud optical depth because transmittance is a concave function of cloud optical depth. In other words, we averaged the natural logarithm of cloud optical depth, and then transformed the average back to obtain the 1 min mean. Additionally, to use the same data set for investigating interdependence of cloud macrophysical, microphysical and optical properties in Sect. 4, we further excluded time periods if the effective radius was outside the range between 3 and 100 µm. This exclusion process led to a final sample size of 5200 min of data points during the period 2005–2007 that represents ~35 % of daytime stratiform cases.

Figure 6 shows histograms of 1 min averaged cloud optical depth and a scatter plot of retrievals from lidar solar background signals against those from flux measurements. Both data sets reveal an occurrence peak at optical depth of 15–20, but an evident discrepancy occurs in the optical depth bin of 0–5. The reason for the lack of small optical depth in lidar-based retrievals is partly because their corresponding LWP values have always been zero or negative and therefore are excluded, implying that it remains challenging for two-channel microwave radiometers to detect very optically thin clouds.

The mean cloud optical depth from lidar measurements is 30, smaller than cloud-mode retrievals by three optical depths. The correlation coefficient is 0.95, while the root-mean-squared difference between the two is 8 (24 % relative to the mean of cloud-mode retrievals).

4 Interdependence of stratiform cloud properties

4.1 Macrophysical properties versus optical depth

Using the same stratiform cloud cases shown above, we investigate how cloud macrophysical and microphysical properties vary with optical depth in non-drizzling and drizzling stratiform clouds, categorised by a reflectivity threshold of −15 dBZ as described in Sect. 2.2. Figure 8a shows that non-drizzling clouds occur more frequently at optical depths of 10–20, while drizzling clouds have a relatively uniform frequency distribution throughout the entire optical depth range. Using an adiabatic cloud model for non-drizzling clouds, Boers and Mitchell (1994) showed that LWP, cloud geometric thickness H and optical depth τ follow LWP ∝ H^2, τ ∝ H^{2/3}, and thus LWP ∝ τ^{6/5}. Not surprisingly, Fig. 8b shows...
that LWP indeed increases approximately linearly with \( \tau \) for both cloud categories. LWP in non-drizzling clouds is proportional to \( \tau^{1.09 \pm 0.01} \) with 95\% confidence intervals; the exponent is slightly smaller than the predicted value of 1.2 under an adiabatic assumption. LWP in drizzling clouds is generally 20–40\% larger than those in non-drizzling clouds.

Similar to LWP, Fig. 8c shows that \( H \) also increases with \( \tau \). Using 1 min averaged ARM data from these stratiform cloud cases, the relationship between \( H \) (in metres) and \( \tau \) can be approximated by

\[
H = (308 \pm 15) \cdot \tau^{0.25 \pm 0.01} \quad \text{for non-drizzling clouds,} \tag{2}
\]

\[
H = (513 \pm 51) \cdot \tau^{0.16 \pm 0.03} \quad \text{for drizzling clouds,} \tag{3}
\]

corresponding to correlation coefficients of 0.95 and 0.79, respectively. These relationships indicate that the geometric thickness in drizzling clouds is at least 10\% larger than that in non-drizzling clouds at a given \( \tau \). We have also found that these relationships vary little when taking hourly means rather than 1 min averages. Using the adiabatic approximation as explained above, the exponents in non-drizzling and drizzling clouds from ARM data are both much smaller than the predicted value of 0.6.

Cloud geometric thickness derived from Eqs. (2) and (3) is compared to the results for marine stratocumulus off the coast of California during the First ISCCP Regional Experiment. Based on Minnis et al. (1992), their relationship between \( H \) and \( \tau \) can be re-written as the following:

\[
H = 58 \cdot \tau^{0.56}, \tag{4}
\]

where \( H \) was retrieved from hourly averaged surface ceilometer and acoustic sounder measurements; \( \tau \) was estimated from Geostationary Operational Environmental Satellite visible and infrared radiances. The relationships obtained here suggest that the geometric thicknesses in continental stratiform clouds can be thicker than marine stratocumulus by at least 35\% for cloud optical depths less than 80.

4.2 Cloud effective radius versus optical depth

Unlike LWP and \( H \), Fig. 8d shows that cloud effective radius has a different dependence on optical depth between non-drizzling and drizzling clouds. The strong positive correlation of 0.8 between cloud effective radius and optical depth in non-drizzling clouds is consistent with many studies using airborne and spaceborne remote sensing measurements (e.g. Han et al., 1994; Nakajima and Nakajima, 1995; Harshvardhan et al., 2002), but the asymptotic radius from the ARM data is \( \sim 10\mu m \), smaller than the so-called critical radius (\( \sim 15\mu m \)) reported in literature for marine low clouds (Nakajima and Nakajima, 1995; Kobayashi and Masuda, 2008; Painemal and Zuidema, 2011). Additionally, these non-drizzling clouds show \( r_{\text{eff}} \) proportional to \( \tau^{0.11 \pm 0.01} \).

The exponent of 0.11 is smaller than the value of 0.2 derived from Eq. (4).
from satellite and aircraft measurements for the eastern Pacific stratocumulus (Szczodrak et al., 2001), and the theoretical value of 0.2 derived under the assumption of adiabatic and constant cloud droplet number concentration $N_d$ (Lohmann et al., 2000), meaning that the condition at the ARM Oklahoma site may be slightly sub-adiabatic, and/or that $N_d$ variation with height is not negligible.

For drizzling clouds, Nakajima and Nakajima (1995) showed that cloud effective radius decreased from 20 to 10 µm with an increase in $\tau$ from 5 to 20. Similarly, our result shows a negative correlation ($-0.75$) with a 99 % confidence level for drizzling clouds. The negative correlations between cloud effective radius and optical depth in drizzling clouds can be explained by precipitation influence, which possibly reduces cloud optical depth through the removal of droplets (Boers and Rotstayn, 2001). Further analyses reveal that a number of drizzling clouds with small optical depths indeed have large effective radii greater than 50 µm, often found at the end of a precipitation system passing over. These cases, however, occurred less frequently compared to those with small effective radii, resulting in the mean cloud effective radii fluctuating between 12 and 18 µm with large standard errors at small optical depths.

Since the correlation between $\tau$ and $r_{eff}$ is positive in non-drizzling clouds but negative in drizzling clouds, the difference in $r_{eff}$ between two types of clouds decreases with increasing cloud optical depth, which is a result of Fig. 8b. Across all optical depth bins, Fig. 8b shows that LWP in drizzling clouds is consistently ~85 g m$^{-2}$ larger than that in non-drizzling clouds. Compared to cases with small $\tau$, this extra LWP in drizzling clouds distributes to more droplets in cases with large $\tau$, leading to a smaller increase in $r_{eff}$ (as shown in Eq. (1) having a denominator $\tau$).

In short, the difference between non-drizzling and drizzling clouds at a given cloud optical depth mainly ranges between 2 and 7 µm with a mean of 5 µm (Fig. 8d), similar to the finding from satellite observations in marine stratocumulus (Kubar et al., 2009). This mean size difference between two cloud categories is clear in Fig. 9a, showing that the distribution of non-drizzling clouds peaks at 6–8 µm, with a mean of 8 µm, and the distribution of drizzling clouds peaks at 10–12 µm, with a mean of 13 µm.

4.3 Interdependence derived from the ARM Min product

To examine whether this interdependence is sensitive to the choice of cloud optical depth product, we repeated the same analysis using the ARM Min product as shown in Fig. 10. Firstly, similar to results derived from lidar-based retrievals, LWP in non-drizzling clouds is proportional to $\tau^{1.01\pm0.01}$ with 95 % confidence intervals. LWP in drizzling clouds are also larger than those in non-drizzling clouds, although the difference between the two varies in a larger range between 30 and 150 g m$^{-2}$. Secondly, cloud geometric thickness can be approximated by

$$H = (249 \pm 12) \cdot \tau^{0.30\pm0.01}$$

for non-drizzling clouds, and

$$H = (447 \pm 33) \cdot \tau^{0.20\pm0.02}$$

for drizzling clouds.
Although the exponents and prefactors of Eqs. (5) and (6) are different from Eqs. (2) and (3), the geometric thicknesses derived from the ARM Min and lidar-based retrievals are similar, as shown in Fig. 10c. Finally, while the negative correlation between \( r \) and \( r_{\text{eff}} \) in drizzling clouds holds in Fig. 10d, the more robust positive correlation in non-drizzling clouds, found in Fig. 8d and satellite observations, disappears due to a relatively flat \( r_{\text{eff}} \) of \( \sim 8 \mu m \) across all optical depth bins.

In short, the relationships of LWP and \( H \) with \( r \) are not sensitive to the choice of cloud optical depth product, but this is not true for the correlation between \( r \) and \( r_{\text{eff}} \). This highlights the potential importance of having comparable FOVs among various instruments for investigating \( r-r_{\text{eff}} \) correlation, although properly addressing this issue may require more thorough simulation experiments.

### 4.4 Implication on drizzle delineation

Taking a different view, now we use the same data set as shown in Fig. 8 to investigate how LWP, \( H \) and \( r \) vary with \( r_{\text{eff}} \). Figure 9b–d shows that properties between non-drizzling and drizzling clouds differ the most in the \( r_{\text{eff}} \) range of 7–11 \( \mu m \), although this could be a result of a relatively smaller sample size outside this \( r_{\text{eff}} \) range. Specifically, Fig. 9d shows that optical depth of non-drizzling clouds increases with \( r_{\text{eff}} \) and changes little at \( r_{\text{eff}} \) beyond 7 \( \mu m \). The relatively small change in \( r \) is also found in the \( r_{\text{eff}} \) range of 7–15 \( \mu m \) for drizzling clouds; this is similar to the finding in satellite observations (Kobayashi and Masuda, 2008), but their data showed such behaviour only when \( r_{\text{eff}} \) was larger than a critical value of \( \sim 15 \mu m \). Since Kobayashi and Masuda (2008) used 21 day measurements from the Tropical Rainfall Measuring Mission satellite and sampled tropical marine warm clouds, the difference in the critical effective radius (7 vs. 15 \( \mu m \)) may be due to the regional variability of precipitating clouds. Additionally, the definition of this critical effective radius is rather loose, and its value can strongly depend on how and at which altitude cloud effective radii were estimated. The difference in the resulting critical value of effective radius between airborne/spaceborne measurements and the ARM data can be partly due to the fact that retrievals from the former is mainly determined by droplets at cloud tops, while the latter is determined by the entire cloud layer (Platnick, 2000; Chiu et al., 2012).

Results from Figs. 8d and 9d imply that it is plausible to delineate drizzling clouds using a simple threshold; for example, we can roughly classify clouds as drizzling when cloud effective radius exceeds a critical value \( r^* \) of 10 \( \mu m \) (Fig. 8d) or when cloud optical depth exceeds 40 (Fig. 9d). Similarly, based on satellite retrievals and ground-based radar measurements, Nauss and Kokhanovsky (2006) proposed a more sophisticated delineation function, given as

\[
r^* = \frac{A}{\tau},
\]

Table 1. Contingency table used to evaluate drizzle delineation methods. A–D represent the number of hits, false alarms, misses and correct negatives, respectively.

<table>
<thead>
<tr>
<th>Reference observations</th>
<th>New method</th>
<th>Yes</th>
<th>No</th>
</tr>
</thead>
<tbody>
<tr>
<td>Yes</td>
<td>A</td>
<td>B</td>
<td></td>
</tr>
<tr>
<td>No</td>
<td>C</td>
<td>D</td>
<td></td>
</tr>
</tbody>
</table>

where coefficient \( A \) is 920 \( \mu m \) and the critical value \( r^* \) varies with cloud optical depth \( \tau \). To evaluate how well these methods discriminate between non-drizzling and drizzling clouds (i.e. a binary classification), we computed the Heidke skill score (HSS) from a contingency table (Table 1), defined as

\[
\text{HSS} = \frac{2(AD - BC)}{(A+C)(C+D) + (A+B)(B+D)}. \tag{8}
\]

HSS not only measures the proportion of correct classifications (including both correct hits and negatives), but more importantly, also takes into account the expected skill obtained by chance in the absence of any skill (Barnston, 1992). In general, an HSS of 0 indicates no skill, while 1 represents perfect skill.

Figure 11 summarises HSS using three different methods. Firstly, using a simple fixed cloud effective radius as the delineation threshold (red lines), the optimal threshold that maximises HSS is 10 \( \mu m \), agreeing with results in Fig. 8d. Secondly, applying a fixed threshold of cloud optical depth (blue lines), the optimal threshold is \( \sim 42 \) and HSS is similar in the optical depth range between 40 and 45. Note that the maximum of HSS using the optimal optical depth threshold is not as good as that from an effective radius threshold of 10 \( \mu m \). Thirdly, a dynamic threshold of cloud effective radius derived by Eq. (7) apparently yields a higher HSS (0.52), compared to the previous two simple methods; the optimal coefficient \( A \) is 380 \( \mu m \), rather than 920 \( \mu m \) found in satellite observations (Nauss and Kokhanovsky, 2006) for convective systems over central Europe taken during the extreme summer floods in 2002. It is expected that the coefficient \( A \) varies with cloud type, site location, and more importantly, the threshold of rain rate used to define drizzle (\( \sim 0.006 \text{ mm h}^{-1} \) in our cases).

Since HSS is dependent on the frequency of occurrence of an event, we further test our delineation thresholds using the symmetric extremal dependence index (SEDI), which is independent of occurrence frequency and thus works for both common and rare events (Ferro and Stephenson, 2011). SEDI is defined as

\[
\text{SEDI} = \frac{\ln F - \ln H + \ln(1 - H) - \ln(1 - F)}{\ln F + \ln H + \ln(1 - H) + \ln(1 - F)}, \tag{9}
\]

where

\[
H = \frac{A}{A + C} \quad \text{and} \quad F = \frac{B}{B + D}.
\]
Similar to HSS, a SEDI value of 0 indicates no skill, while 1 represents perfect skill. As Fig. 12 shows, the optimal cloud effective radius and cloud optical depth thresholds are $\sim 10 \mu m$ and $\sim 40$, respectively. When considering a dynamic threshold, the optimal coefficient $A$ of 340 $\mu m$ is found. Overall, the optimal thresholds from SEDI are similar to those derived from HSS.

In short, depending on the availability of measurements, one can use a cloud optical depth of 40 as the simplest way for drizzle delineation in the absence of LWP and radar measurements, although this threshold may depend on ambient aerosol loading. If co-incident LWP measurements are available, the dynamic threshold of cloud effective radius given in Eq. (7) with a coefficient A of 340–380 $\mu m$ is a better approach to delineating drizzle for mid-latitude continental stratiform clouds.

5 Summary

To better represent clouds in weather and climate models, long-term global measurements can provide direct constraints and improve our knowledge of cloud and precipitation formation, and their interactions with radiation and aerosol. In particular, low warm clouds strongly influence global climate through their impacts on Earth’s radiation and water energy cycle. While marine low clouds have been extensively studied, continental warm clouds received relatively little attention partly due to the fact that the majority of satellite retrievals work best over oceans.

Using ground-based measurements at the ARM Oklahoma site during the period 2005–2007, we conducted an extensive analysis for mid-latitude continental low-level clouds. To retrieve cloud optical depth, we developed a novel method that capitalised on unexploited solar background light that is currently treated as noise and has largely inhibited lidar applications in all-sky conditions and during daytime. This new technique works well; when compared to other benchmarks, the mean bias of cloud optical depth is around 2 and the root-mean-squared difference is 8 (24 % relative to the mean optical depth). Since lidars have a FOV much closer to those of microwave radiometers than conventional hemispheric-viewing radiometers, it is more appealing to use lidar-based cloud retrievals to understand the linkage between cloud macrophysical, microphysical and optical properties.

A number of key features are found in the relationships between LWP, geometric thickness $H$, droplet effective radius $r_{eff}$ and cloud optical depth $\tau$. Firstly, LWP and $H$ follow a power-law relationship with positive exponents with $\tau$; LWP and $H$ in drizzling clouds are generally 20–40% and at least 10% higher than those in non-drizzling clouds, respectively. Similar to LWP, $r_{eff}$ also increases with $\tau$ following a power law for non-drizzling clouds, but this does not hold for drizzling clouds. In the presence of drizzle, a negative correlation is found between $r_{eff}$ and $\tau$; $r_{eff}$ also tends to be 5 $\mu m$ larger than droplet sizes in non-drizzling clouds.

While several aircraft and satellite observations have suggested that $r_{eff}$ of the order of 15 $\mu m$ may be a good indicator to distinguish between non-drizzling and drizzling marine clouds, we found that a threshold of $\sim 10 \mu m$ works better...
for ground-based observations. The difference in threshold between various observational platforms is likely attributed to the fact that satellite retrievals are mainly determined by properties at cloud tops and, conversely, ground-based retrievals utilise the full cloud profile. If co-incident LWP measurements are available, a dynamic threshold of cloud effective radius given in Eq. (7) with a coefficient $A$ of 340–380µm is a better approach to delineating drizzle for mid-latitude continental stratiform clouds.

We have demonstrated a novel retrieval method using untapped solar background signals in lidar measurements, which greatly extends lidar applications from cirrus to all types of clouds, and provides a new approach to distinguishing between non-drizzling and drizzling clouds. With collocated radar and LWP measurements, the new retrieval can also be used to compare and contrast drizzle and drizzle-free cloud properties. This new method can be easily adapted to the exiting lidar networks if collocated sun photometer measurements are available, including the high-density ceilometer networks in the United Kingdom, France and Germany that have been established for monitoring volcanic plumes (Heese et al., 2010). Combined with the ability of lidars to resolve vertical distributions of aerosol properties below cloud layers, collocated and simultaneous measurements of aerosol and cloud are also possible, which can help advance our understanding of aerosol–cloud interactions.

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