On the nature and extent of optically thin marine low clouds


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1 Macrophysical properties of optically thin marine low clouds over the nonpolar oceans (60°S–60°N) are measured using 2 years of full-resolution nighttime data from the Cloud-Aerosol Lidar with Orthogonal Polarization (CALIOP). Optically thin clouds, defined as the subset of marine low clouds that do not fully attenuate the lidar signal, comprise almost half of the low clouds over the marine domain. Regionally, the fraction of low clouds that are optically thin (f_{thin,cld}) exhibits a strong inverse relationship with the low-cloud cover, with maxima in the tropical trades (f_{thin,cld} > 0.8) and minima in regions of persistent marine stratocumulus and in midlatitudes (f_{thin,cld} < 0.3). Domain-wide, a power law fit describes the cloud length distribution, with exponent β = 2.03 ± 0.06 (±95% confidence interval). On average, the fraction of a cloud that is optically thin decreases from ~1 for clouds smaller than 2 km to <0.3 for clouds larger than 30 km. This relationship is found to be independent of region, so that geographical variations in the cloud length distribution explain three quarters of the variance in f_{thin,cld}. Comparing collocated trade cumulus observations from CALIOP and the airborne High Spectral Resolution Lidar reveals that clouds with lengths smaller than are resolvable with CALIOP contribute approximately half of the low clouds in the region sampled. A bounded cascade model is constructed to match the observations from the trades. The model shows that the observed optically thin cloud behavior is consistent with a power law scaling of cloud optical depth and suggests that most optically thin clouds only partially fill the CALIOP footprint.


1. Introduction

The planetary albedo is a quantity determined by the reflection of sunlight by the Earth’s surface and atmosphere. It is a critical determinant of the Earth’s temperature and climate, and it is therefore imperative that its controlling factors be understood and adequately quantified. Without clouds, the planetary albedo would be about 0.15, due to the low reflectance of the Earth’s surface (most of which is ocean). Highly reflective surfaces, such as those characteristic of the polar regions and unvegetated deserts, are notable exceptions but cover a small part of the planet and do not contribute strongly to the average surface albedo. However, the average albedo of the Earth (~0.3) is roughly doubled by the presence of clouds [Harrison et al., 1990], and low clouds are the dominant type by contribution to albedo [Hartmann et al., 1992].

[2] Among albedo-influencing components of the Earth system, clouds are unique in both their extreme transience and in their remarkable range of sizes and forms. Although we have a reasonably accurate assessment of the global impact of clouds on the planetary albedo, our understanding of how clouds of different sizes and types contribute to the albedo remains rather poor. Although it is known that low clouds are the dominant contributor to the albedo globally [Hartmann et al., 1992], it is uncertain what the relative contributions of the different low-cloud types (cumulus, stratocumulus, stratus) are to the overall albedo of low clouds.

[3] Moist convection over the oceans has a varied morphology. Marine low clouds range from expansive stratocumulus decks stretching hundreds of kilometers, to small trade wind cumulus a few tens to hundreds of meters in extent [Hozumi et al., 1982; Wood and Field, 2011]. We are currently unable to quantify how different low-cloud systems distribute condensate horizontally and vertically to determine their albedo. This is partly because we do not fully understand how albedo scales with cloud cover, as the latter is frequently difficult to measure with the moderate-resolution (~1 km) passive sensors used to generate our cloud cover climatologies [Zhao and Di Girolamo, 2006; Jones et al., 2012]. We still do not have proven measures of cloud condensate amounts, particularly for broken and optically thin
cloud fields [Turner et al., 2007]. Many questions are therefore still largely unanswered. For example, per unit of cloud cover, are trade cumulus clouds optically thinner than strato-cumulus? What fraction of low clouds is optically thin and how are these optically thin clouds partitioned by cloud size and type? What resolution do our passive sensors need to have in order to accurately determine cloud cover? [5] Satellite data show that the albedo of extensive marine low clouds is quite low [Bender et al., 2011], and that their optical thickness is highly variable [e.g., Rossow et al., 2002]. It is reasonable to infer that a significant fraction of them must be optically thin. These optically thin clouds are not currently well resolved observationally, but are globally pervasive and potentially important for climate studies [Garrett and Zhao, 2006; Turner et al., 2007; Zuidema et al., 2012]. For this work, we define operationally an optically thin cloud as a cloud that does not fully attenuate a lidar backscatter signal. This corresponds to unbroken clouds with optical depth (τ_{cld}) less than ~3. Analysis of τ_{cld} from Moderate Resolution Imaging Spectroradiometer (MODIS) Level 3 liquid cloud retrievals over an ocean region spanning ±45° latitude suggests that clouds with τ_{cld} < 3 comprise ~30% of marine low clouds, and contribute perhaps 15% of their albedo (Figure 1). But since many of these clouds are known to be smaller than the resolution of the MODIS pixels [e.g., Zhao and Di Girolamo, 2007], it is important to appreciate that these estimates are likely to be a function of sensor resolution. How does the picture change when clouds are viewed with an active sensor with a much smaller footprint size? Indeed, does the use of metrics such as cloud cover even make sense if a significant fraction of the clouds have very low optical thicknesses?

Figure 1. Cumulative fraction of MODIS L3 liquid water cloud optical depth τ_{cld} for 12 months of ocean daytime data spanning ±45° latitude. Cloud albedo is approximated as τ_{cld}/τ_{cld} + 7.7 [Lacis and Hansen, 1974]. MODIS data suggest approximately one third of low clouds over the oceans have optical depth less than 3 (optically thin cloud definition, this study). In addition, such clouds may contribute up to 15% to total low-cloud albedo. The inset describes relative scales of the CALIOP receiver footprint at single-shot resolution and MODIS pixel at cloud product resolution. It should be noted that at most three CALIPSO footprints will traverse a given MODIS (on Aqua) pixel.

Optically thin and broken clouds are not only important because of their contribution to the planetary albedo, but also because aerosol retrievals are problematic where clear regions are in close proximity to clouds and contamination of clear pixels by subpixel-scale optically thin clouds is common [e.g., Zhang and Reid, 2005; Charlson et al., 2007; Koren et al., 2007]. A trade wind cumulus study reported that a third of MODIS pixels diagnosed by the cloud mask as being clear actually contained subpixel-scale cloud [Zhao and Di Girolamo, 2006]. Over the global ice-free oceans, 50% of marine low clouds are separated from each other by less than 5 km [Várnai and Marshak, 2011], suggesting that a high fraction of clear sky over ocean is in close proximity to cloud. Other recent studies of the clear-cloud transition region using both in situ [Twyoh et al., 2009] and remote sensing measurements [Su et al., 2008; Tackett and Di Girolamo, 2009; Redemann et al., 2009] suggest that aerosol properties in the vicinity of clouds are different from those 2–5 km from the cloud. Some of this difference may be attributed to aerosol hygroscopic growth in the high relative humidity air surrounding cloud, but the relative magnitudes of cloud contamination and aerosol hygroscopic growth effects on retrieved aerosol properties is not known. Observations show that enhancement of near-cloud reflectances due to undetected or unresolved cloud contamination of clear pixels leads to passive sensor aerosol optical depth retrieval overestimates of 10%–20% [Zhang and Reid, 2005].

[7] Although passive sensors provide excellent coverage, it is fundamentally challenging to identify which atmospheric constituents are producing the measured top of atmosphere (TOA) signals, and at what altitudes they reside. Spaceborne lidar provides a metric of the TOA reflectance, integrated over the specified layers of interest, and offers a closer link to the atmospheric constituent producing that signal. A frequency distribution of such a metric, integrated attenuated backscatter (IABS), derived from high-resolution (90 m telescope footprint) Cloud-Aerosol Lidar with Orthogonal Polarization (CALIOP) is presented in Figure 2. A bimodal distribution of IABS is observed with modes representing clear and cloudy atmospheric states as indicated (Figure 2a). The nonzero minimum between clear and cloudy states, which exemplifies the fundamental challenge in separating clouds from clear sky, has been termed the “continuum” region [Charlson et al., 2007]. With detailed classification with CALIOP’s full-resolution cloud clearing algorithm [Vaughan et al., 2009], spaceborne lidar is able to inform us that this “continuum,” or clear-cloud transition region, is populated almost exclusively with clouds (Figure 2b). These clouds must, by their virtue of having low visible reflectance, be optically thin. IABS can be small either because the cloud is optically thin or because the laser was attenuated fully some distance into an optically thick cloud. The latter case is manifest in Figure 2b as the smaller values of IABS categorized as optically thick clouds. A frequency distribution alone cannot inform us of their spatial structure, but previous work suggests a prevalence of detraining cloud elements, dissipating, or nascent clouds in marine low-cloud systems [Koren et al., 2009]. However, the CALIOP data are showing us that optically thin clouds are not only limited to the continuum region (approximately defined as the IABS region spanning...
5 \times 10^{-2} \text{ sr}^{-1} to 3 \times 10^{-1} \text{ sr}^{-1}), but contribute to TOA reflectance throughout the entire cloud IABS distribution (Figure 2b). Indeed, over nonpolar oceans, we find that almost half of the marine low clouds are optically thin by our definition of surface detectability.

Despite the apparent prevalence of optically thin clouds, very little is currently known about their macrophysical properties (spatial distribution, length, vertical distribution, geometric depth) on a global scale. This study aims to provide insight into these properties, and thereby provide new information on a poorly characterized and understood subset of marine low clouds. Compared to satellite observations, large-scale models tend to overestimate the frequency of optically thick low clouds and underestimate the frequency of optically thin low clouds (e.g., a Community Atmosphere Model study [Kay et al., 2012]). At the low end of the optical depth distribution, model frequencies are close to an order of magnitude too low compared to passive remote sensing satellite observations [Kay et al., 2012]. However, it is important to appreciate that there are major differences between the various observational data sets [see Kay et al., 2012, Figure 3], making accurate assessment of model representation of optically thin clouds difficult.

The primary data set employed for this study is the CALIOP full-resolution cloud mask (60 m vertical resolution and 90 m footprint diameter, spatially sampled every 335 m) at 1064 nm. In addition to providing new near-global observations of the macrophysical properties of optically thin marine low clouds, we also investigate possible sensor resolution detection issues by comparing CALIOP with data from NASA’s airborne High Spectral Resolution Lidar (HSRL). The ensuing sections of this manuscript are ordered as follows. Data set description and analysis methodology are detailed in section 2. Results are presented in section 3. A discussion of the implications is augmented and informed by the use of a simple fractal model in section 4. Our findings are summarized in section 5.

2. Data Set Description and Analysis Method

2.1. CALIOP Data Set

Two years (September 2006 to August 2008) of vertically resolved cloud data from the full-resolution CALIOP Vertical Feature Mask (VFM) Version 3.01 [Vaughan et al., 2009] over the nonpolar oceans (±60° latitude) are used in this study. The VFM provides vertically resolved cloud and aerosol layer information generated by the CALIOP multi-resolution layer identification and scene classification algorithms. Full-resolution layer information is obtained from the high-resolution cloud clearing tool within the layer detection algorithm. In the lowest 4 km of the atmosphere the 1064 nm channel (60 m vertical resolution) is used for full-resolution cloud layer identification, and any feature detected above the surface is classified as a cloud [Vaughan et al.,
This wavelength is less sensitive to aerosol than the 532 nm channel and this improves cloud-aerosol discrimination. A detailed mission overview and comprehensive instrument details are presented by Winker et al. [2009] and Hunt et al. [2009], respectively. CALIOP has high sensitivity at low-cloud optical depths: nighttime minimum detectable particulate extinction at 1064 nm is 0.165 km$^{-1}$ for layers greater than 180 m in depth, equivalent to $\tau_{cd}$ greater than 0.03, and is 1.65 km$^{-1}$ for layers 60–180 m (minimum 60 m), equivalent to $\tau_{cd}$ greater than 0.1 [Chepfer et al., 2012]. CALIOP provides near-global ($\pm8^\circ$ latitude), vertically resolved, atmosphere-only backscatter data, with no contamination from surface return or 3D radiation/cloud adjacency effects, unlike radiances measured from passive sensors [Wen et al., 2007; Várnai and Marshak, 2009]. However, sampling is sparse. The lidar return signal is limited to a 90 m diameter receiver footprint, so no off-track atmospheric information can be provided. Results are only representative of a single sampling time for every location. In addition, consecutive laser footprints are not contiguous; horizontal separation between each footprint is 335 m and consecutive nighttime (and daytime) orbits are $\sim$2600 km apart at the Equator, requiring a temporally extensive data set for representative sampling.

[11] Our focus is optically thin marine low clouds. In this study, a low cloud is defined as having a cloud top height (CTH) less than or equal to 3 km. A high cloud is any cloud higher than this. We define an optically thin cloud as a cloud detected at full resolution that does not fully attenuate the particulate extinction at 1064 nm, equivalent to $\tau_{cd}$ greater than 0.1 [Chepfer et al., 2012]. A high cloud is any cloud with CTH less than or equal to 3 km. A low cloud is defined as having a cloud top height (CTH) less than or equal to 3 km. We define an optically thin cloud as a cloud detected at full resolution that does not fully attenuate the particulate extinction at 1064 nm, equivalent to $\tau_{cd}$ greater than 0.1 [Chepfer et al., 2012]. CALIOP provides near-global ($\pm8^\circ$ latitude), vertically resolved, atmosphere-only backscatter data, with no contamination from surface return or 3D radiation/cloud adjacency effects, unlike radiances measured from passive sensors [Wen et al., 2007; Várnai and Marshak, 2009]. However, sampling is sparse. The lidar return signal is limited to a 90 m diameter receiver footprint, so no off-track atmospheric information can be provided. Results are only representative of a single sampling time for every location. In addition, consecutive laser footprints are not contiguous; horizontal separation between each footprint is 335 m and consecutive nighttime (and daytime) orbits are $\sim$2600 km apart at the Equator, requiring a temporally extensive data set for representative sampling.

[13] To provide an example of optically thin clouds and their spatial context, we present colocated Wide Field Camera (WFC) band-averaged radiance (620–670 nm) for a 10 km orbit segment at 125 m horizontal resolution, and CALIOP full-resolution Level 1 daytime IABS at 532 nm integrated from 0 to 20 km, for a broken cloud field in the northwest subtropical Atlantic (Figure 3). Along with CALIOP, the WFC is part of the Cloud Aerosol Lidar with Integrated Pathfinder Satellite Observation (CALIPSO) satellite’s payload. The WFC context (Figure 3a) suggests that some detected optically thin cloud profiles are cloud edges that partially fill the lidar field of view. We explore this issue further in section 4. In general, peaks and troughs observed in both the radiance and IABS data track each other remarkably well (Figure 3b). Optically thick cloud profiles (fully attenuated profiles) tend to be associated with stronger visible radiances. However, radiances associated with optically thick clouds at $\sim$6 km along the orbit segment are similar to some of the radiances for optically thin cloud profiles elsewhere. The lack of a complete correspondence between radiance and IABS is indicative of the different nature of the lidar backscatter signal and the visible reflectance. Multiple scattering affects the signals in different ways. In addition, it is likely that beam filling in this broken cloud field affects visible radiance and lidar signals differently. The clear evidence in this example for small and often optically thin cloud elements on the scale of the lidar footprint compels us to understand how optically thin clouds are related to cloud horizontal scales globally and regionally.

2.2. HSRL Data Set

[14] In addition to a near-global examination of CALIOP’s view of optically thin marine low clouds, we also compare colocated CALIOP cloud fractions and cloud lengths with those estimated using the airborne NASA LaRC High Spectral Resolution Lidar (HSRL) [Hair et al., 2008]. Although HSRL is typically employed for aerosol studies and/or instrument validation [e.g., Kacenelenbogen et al., 2011; Rogers et al., 2011] for this work we use derived HSRL altitudes of the ocean surface and cloud top. Cloud data for the comparison are drawn from four daytime and four nighttime spatially and temporally matched HSRL underflights of the CALIOP orbit track over the tropical and subtropical western Atlantic (Table 1), and we include only data for flight segments with aircraft altitude greater than 8 km. HSRL cloud screening is based upon a wavelet transform technique [Su et al., 2008]. HSRL and CALIOP do not have the same footprint dimensions due to different sampling configurations. The HSRL sample area is composed of 60 m along-track averages of contiguous 8 m single-shot footprints, yielding a sample area 8 m $\times$ 60 m in the across-track and along-track directions, respectively. The 60 m along-track dimension results from 2 Hz sampling at aircraft ground speed of 120 m s$^{-1}$ (mean value for data analyzed herein).
However, CALIOP 90 m diameter footprints are horizontally separated by 335 m in the along-track direction. Henceforth, we refer to an instrument footprint, pixel size or sample area as a field of view (FOV) for clarity of notation. HSRL data have 30 m vertical resolution. In keeping with the CALIOP analysis, we identify cloudy HSRL profiles as those with CTH less than or equal to 3 km. We use the HSRL ocean surface altitude parameter to determine whether a surface return was detected and thereby discriminate between low and high optical depth clouds, i.e., whether or not the signal was fully attenuated by the cloud before reaching the surface. For the ocean data used, we expect the surface altitude for a cloud profile to be ~0 m if the signal is not fully attenuated. However, the surface altitude detection algorithm is not fully developed (J. Hair, HSRL data summary, unpublished material, 2009) and the detected surface altitude departs somewhat from zero, even for cloud-free profiles over the ocean. Therefore, we define a cloud profile with surface altitude less than or equal to 85 m as optically thin. The upper limit of 85 m is the mean surface altitude plus three standard deviations observed for cloud-free HSRL profiles over the ocean. For the HSRL–CALIOP comparisons, we include only HSRL and CALIOP data that achieve temporal coincidence within 15 min. Both

![Image](orbit_segment.png)

**Figure 3.** (a) A two-dimensional (5 km × 10 km) and (b) one-dimensional (10 km) view of a high-cloud-screened broken marine low-cloud scene, generated from CALIOP data along a 10 km daytime orbit segment over the northwest tropical Atlantic (32°N, 77°W) on 27 May 2007: Figure 3a shows Wide Field Camera (WFC) band-averaged radiance (620–670 nm) at 125 m horizontal resolution with CALIOP ground track position superimposed. The WFC data are continuous, along-track radiances. Symbols along the CALIOP track indicate clear (yellow square), optically thin cloud (white circles), and optically thick clouds (black circles). Figure 3b shows collocated WFC radiance (black line) and Level 1 532 nm attenuated backscatter integrated from 0 to 20 km (IABS; pink line) for the same segment as in Figure 3a. Symbols identifying clear profiles and cloud profiles flagged as optically thin or optically thick are also indicated.

<table>
<thead>
<tr>
<th>Date</th>
<th>Location</th>
<th>Overpass Time</th>
<th>Cloud Type</th>
</tr>
</thead>
<tbody>
<tr>
<td>26 Jan 2007</td>
<td>36.88° N, 75.52° W</td>
<td>07:11</td>
<td>Cumulus</td>
</tr>
<tr>
<td>23 May 2007</td>
<td>32.41° N, 77.07° W</td>
<td>18:32</td>
<td>Cumulus</td>
</tr>
<tr>
<td>26 Jan 2008</td>
<td>12.43° N, 60.26° W</td>
<td>17:36</td>
<td>Cumulus</td>
</tr>
<tr>
<td>11 Aug 2010</td>
<td>33.59° N, 65.09° W</td>
<td>17:43</td>
<td>Cumulus/stratocumulus</td>
</tr>
<tr>
<td>18 Aug 2010</td>
<td>22.38° N, 63.95° W</td>
<td>17:48</td>
<td>Cumulus</td>
</tr>
<tr>
<td>22 Aug 2010</td>
<td>13.71° N, 69.04° W</td>
<td>06:39</td>
<td>Cumulus</td>
</tr>
<tr>
<td>24 Aug 2010</td>
<td>21.76° N, 64.09° W</td>
<td>06:15</td>
<td>Cumulus</td>
</tr>
<tr>
<td>26 Aug 2010</td>
<td>20.67° N, 61.25° W</td>
<td>06:02</td>
<td>Cumulus</td>
</tr>
</tbody>
</table>

aTransect lengths range from 34 to 215 km, with an 8 day mean of 126 km.
bSegment midpoint (latitude, longitude).
cTimes are UTC.
dCloud type information from inspection of MODIS Aqua L1B Granule Images (daytime only, http://modis-atmos.gsfc.nasa.gov/IMAGES/index_myd021km.html).
data sets are high cloud screened using the CALIPSO VFM described in section 2.1.

2.3. Analysis Methods

[15] After screening for high cloud, CALIOP full-resolution data are binned into $5^\circ \times 5^\circ$ latitude-longitude grid boxes. Gridding the data allows us to examine the geographical variations of the various cloud features observed. We define a transect as being the portion of ground track for each CALIPO orbit within a grid box. Separate analysis methods are applied to these binned data to produce cloud cover statistics, and cloud length distributions.

2.3.1. Cloud Cover Analysis Method

[16] Monthly cloud cover values calculated from counts of full-resolution clear profiles and low-cloud profiles within a grid box (Table 2) are averaged to produce 2 year mean low-cloud cover values for each grid box. Domain-wide cloud cover is the mean of all grid box values.

[17] Minima in high-cloud-screened CALIOP profile counts (Figure 4a) exhibit a similar pattern to ISCCP–D2 cloud amount maxima for mid plus high clouds (available from http://isccp.giss.nasa.gov/products/browsed2.html). ISCCP annual mean data (not shown) indicate maximum high cloud amounts in the western equatorial Pacific and northern Indian Ocean (amounts $>40\%$), and midlevel cloud amount maxima (30–40%) in the Northern and Southern Hemisphere storm track regions.

2.3.2. Cloud Length Distribution Method

[18] Cloud length is calculated for individual cloud entities. We define a cloud entity as a series of consecutive marine low-cloud profiles, bounded on each end by a clear profile. The cloud length data are binned into logarithmically spaced size bins, with the lower size limit set by instrument resolution, and the upper limit defined by the sampling method discussed below. An additional criterion is applied to the size distribution data that is not required for cloud cover analysis: we require a transect to comprise at least 100 km of consecutive high-cloud-screened profiles, thus ensuring a long enough transect length to capture most of the contributions to cloud cover from clouds of different sizes. A transect may be less than 100 km and therefore excluded from the cloud length data set because (a) it “clips” the edge of a $5^\circ \times 5^\circ$ grid box, (b) high-cloud screening eliminates a portion of the transect so that the remaining segment lengths are shorter than 100 km, or (c) a portion of transect is land. However, even transects longer than 100 km have a maximum length ($\sim550$ km) imposed by the $5^\circ \times 5^\circ$ gridding. For every cloud length calculated we retain the length of the transect containing the observed cloud, and use this information to formulate a simple correction to the size distribution (see Wood and Field [2011] for details) to account for this upper limit cutoff as follows:

$$n_{\text{corr}}(L)\,dL = n(L)\,dL\frac{L_{\text{tran}}}{(L_{\text{tran}} - L_{\text{min}})}$$

where $n_{\text{corr}}(L)$ is the corrected cloud count, $L_{\text{tran}}$ is the mean of the binned transect lengths with values ranging from 100 km to $\sim550$ km, $L_{\text{min}}$ is the bin minimum cloud length for each size bin, and $n(L)\,dL$ is the number of clouds sampled with sizes between $L$ and $L + dL$. As $L_{\text{min}}$ approaches $L_{\text{tran}}$, the magnitude of the correction to $n(L)\,dL$ increases. To help elucidate the contributions of optically thin and optically thick profiles to clouds of a given size, we construct size distributions for “majority optically thick” entities (contiguous cloudy profiles for which over 90% of contributing profiles are optically thick) and “majority optically thin” entities (over 90% of contributing profiles are optically thin). The threshold of 90% was chosen to accommodate observed variability in cloud optical depth [Wood and Taylor, 2001; Rossow et al., 2002], while ensuring that the clouds analyzed were predominantly optically thick or optically thin, respectively. The cloud length data set is a subset of the data set described in Section 2.3.1, and is composed of approximately 3 times fewer profiles (see Figures 4a and 4b) because of the minimum transect length restriction.

3. Results

3.1. Cloud Cover

[19] High-cloud-screened marine low-cloud cover ($f_{\text{clld}}$) is at a minimum throughout the deep tropics, and is at a maximum in the northern hemisphere and southern hemisphere storm track regions, with additional maxima in the stratocumulus regions on the eastern side of the oceanic subtropical highs (Figure 5a). Values range from 0.10 to greater than 0.80, with a 2 year, domain mean of 0.50. The spatial pattern of cloud cover is similar to climatologies from surface observations (S. G. Warren et al., 2010, Climatic atlas of clouds over land and ocean, http://www.atmos.washington.edu/CloudMap) and remote sensing [Rossow and Schiffer, 1999]. Optically thin low clouds are present in 10–30% of high-cloud-screened profiles over the nonpolar oceans, with no strong geographic variation ($f_{\text{thin}}$; Figure 5b). When expressed as the fraction of marine low-cloud profiles that are optically thin ($f_{\text{thin,clld}}$), a clear spatial pattern emerges, with maxima in the trade wind cumulus regions, and minima in persistent stratocumulus and storm track regions (Figure 5c). The 2 year mean value of
Thin, cold is 0.45 demonstrating that almost half of all marine low clouds are optically thin. While 0.45 is higher than that from MODIS (Figure 1), both sensors indicate a major role for optically thin clouds in determining global cloud cover.

A remarkably strong negative correlation between \( f_{\text{thin,cld}} \) and \( f_{\text{cld}} \) exists at both monthly and annual timescales (Figure 6). In the trade wind regions, where marine low-cloud cover is lower than 0.25 (Figure 5a) [Medeiros et al., 2010], it is remarkable that consistently greater than 80% of the observed clouds are optically thin. In regions typical of marine stratocumulus, where nighttime marine low-cloud cover exceeds 0.80, \( f_{\text{thin,cld}} \) is consistently lower, with values of 0.30 or less being typical. The presence of a large amount of marine low cloud that is optically thin on the scale of the relatively small CALIOP footprint is surprising and warrants further investigation. We devote much of this paper to analyzing these optically thin clouds. However, we should note that this result is consistent with the TOA albedo from Clouds and the Earth’s Radiant Energy System (CERES) data (available from http://eos.atmos.washington.edu/cgi-bin/ceres/disp.pl?ceres.alb.ann.d) of the trade wind regions being quite low (typically 0.16) despite cloud coverage in the trade wind regions of 0.20–0.25. Given observed trade wind TOA albedo and cloud cover, and using a simple cloud albedo – cloud optical depth relation [Lacis and Hansen, 1974] and a simple TOA albedo model (see section 4), we estimate that cloud optical depth \( \tau_{\text{cld}} \) would need to be 2.7–3.0 to produce the observed TOA albedo, if all clouds have the same \( \tau_{\text{cld}} \). While the prevalence of optically thin clouds may appear surprising, it is consistent with regional albedo observations.
3.2. Cloud Length Distribution

Here, we examine the properties of marine low clouds, and especially optically thin clouds, as a function of their horizontal scale. Results presented in this section are derived from analysis of cloud entities. We find that cloud length distributions approximately follow a power law fit \( n(L) = aL^{-\beta} \), where \( n(L) \) is the number of clouds of size \( L \) to \( L + dL \), \( L \) is the bin center, and \( a \) and \( \beta \) are constants, consistent with previous studies [Zhao and Di Girolamo, 2007; Wood and Field, 2011; Benner and Curry, 1998]. For \( \beta = 2 \), each logarithmic size interval (e.g., 100 m to 1 km, 1–10 km, etc.) contributes equally to total cloud cover and \( \beta > 2 \) (<2).

Figure 5. Two year mean ocean nighttime (a) marine low-cloud cover, (b) optically thin marine low-cloud cover, and (c) optically thin fraction of marine low cloud (see Table 2 for fraction definitions). Data are 5° × 5° resolution. Corresponding domain-averaged values are 0.50 (0.25), 0.23 (0.09), and 0.45 (0.28) for the mean (standard deviation) in Figures 5a–5c, respectively. For this calculation, grid box values are weighted by the fraction of each grid box that is ocean, and \( f_{\text{thin, cld}} \) values are also weighted by \( f_{\text{cld}} \).
implies that the distribution is weighted toward smaller (larger) clouds \cite{Wood and Field, 2011}. For the CALIOP data set comprising all low clouds, the size distribution is well fitted by a single power law with $\beta = 2.03 \pm 0.06$ (error at $2\sigma$ level). This is higher than the value of $1.66 \pm 0.04$ reported by \cite{Wood and Field, 2011} for a near-global cloud data set comprising aircraft and satellite data. One explanation may be differences in cloud sampling at the low and high ends of the size distribution. Although \cite{Wood and Field, 2011} aircraft data have a higher resolution than CALIOP, $\sim 100$ m compared to $335$ m, respectively, the aircraft did not sample regions where small cumulus is the dominant cloud type \cite{Wood and Field, 2000, 2011}. Further, CALIOP may undersample large clouds as the upper cloud length limit for this study is $\sim 550$ km compared to $\sim 4000$ km in \cite{Wood and Field, 2011}. Thus, our data set tends to sample smaller clouds on average, and smaller clouds tend to have larger exponents (see below and Figure 12 in \cite{Wood and Field, 2011}).

“Majority optically thin” clouds (defined in section 2.3.2) constitute nearly all of the clouds smaller than 4 km, whereas clouds larger than 35 km have a greater contribution from optically thick profiles (Figure 7). This highlights the increasing importance of optically thin clouds as clouds become smaller.

To further explore differences in cloud length distributions, we focus on three specific cloud regimes (Table 3): (1) a region of persistent stratocumulus off the California coast, (2) a tropical Pacific stratocumulus to cumulus (Sc-Cu) transition region, and (3) a trade wind cumulus region of the tropical Pacific. Cloud length distributions for the regions of stratocumulus and Sc-Cu transition follow a $\beta < 2$ power law fit with values of $1.81 \pm 0.08$ and $1.68 \pm 0.09$, respectively, indicating that large clouds dominate cloud cover (Figure 7, inset). As expected, the trade wind cloud length distribution has a $\beta$ greater than 2 ($2.49 \pm 0.25$) and is dominated by clouds with lengths less than 2 km, consistent with prior studies of trade wind cumulus \cite{Benner and Curry, 1998, Zhao and Di Girolamo, 2007}. Since trade wind regions consist largely of optically thin clouds (Figure 5c), these results paint an emerging picture of small marine low clouds being optically thin. This has implications for the remote sensing of cloud and aerosol properties given current sensor resolution for most passive satellite instruments used to construct cloud climatologies.

### 3.2.1. Cloud Length at Median Cloud Cover

We define a cloud length at median cloud cover ($L_{50}$) for each of the three cloud categories (marine low cloud, majority optically thin cloud, majority optically thick cloud) such that clouds with lengths up to $L_{50}$ contribute 50% to cloud cover for that category. Regions with the smallest coverage of low cloud are associated with smaller $L_{50}$ values (see Figures 5a and 8a), with $L_{50} < 2$ km in trade wind cumulus regions and $L_{50} > 30$ km in stratocumulus regions. Our results for the trades are qualitatively consistent with a recent study in the tropical western Atlantic trade wind region reporting $L_{50}$ of 2 km using high-resolution passive satellite imagery.

**Figure 6.** Joint distribution of monthly nighttime gridded optically thin fraction of marine low cloud as function of marine low-cloud cover overlaid with 2 year mean grid box values (gray). Both monthly and 2 year averaged data exhibit a strong negative correlation. The correlation coefficient for the monthly data is $-0.83$, significant at the 95% confidence level.

**Figure 7.** Cloud length distribution based upon 2 years of gridded data for marine low cloud, majority optically thin cloud, and majority optically thick cloud. Each majority category includes only clouds comprising greater than 90% cloud profiles belonging to either the optically thin or optically thick category. Size bins are logarithmically spaced, and the ordinate axis is normalized frequency. The $\beta = 2$ line is shown for reference. The inset shows marine low-cloud length distribution for the following regions (region center latitude, longitude) California stratocumulus region (Sc; 25°N, 125°W), tropical Pacific stratocumulus to cumulus transition (Sc-Cu Trans; 15°N, 135°W), and tropical Pacific trade wind cumulus (Cu; 15°S, 155°W). Abscissa values are cloud length bin centers.
Table 3. Stratocumulus (Sc), Cumulus (Cu), and Transition From Sc to Cu Regime (Sc-Cu) Region Details: Location, Low-Cloud Cover fcld, and Optically Thin Fraction of Low Cloud fcld,thin

<table>
<thead>
<tr>
<th>Cloud Type</th>
<th>Latitude</th>
<th>Longitude</th>
<th>fcld</th>
<th>fcld,thin</th>
</tr>
</thead>
<tbody>
<tr>
<td>Sc</td>
<td>20°N–30°N</td>
<td>120°W–130°W</td>
<td>0.79</td>
<td>0.36</td>
</tr>
<tr>
<td>Sc-Cu</td>
<td>10°N–15°N</td>
<td>130°W–140°W</td>
<td>0.70</td>
<td>0.37</td>
</tr>
<tr>
<td>Cu</td>
<td>10°S–20°S</td>
<td>150°W–160°W</td>
<td>0.26</td>
<td>0.84</td>
</tr>
</tbody>
</table>

*Here fcld and fcld,thin are 2 year means of monthly values. N = 96 (24 months × 4 grid box values). Also, fcld and fcld,thin, sampling uncertainties are 0.02–0.04 (95% confidence level) and are estimated from the bootstrap method [Efron and Gong, 1983].

visible satellite data [Zhao and Di Girolamo, 2007]. Additionally, we observe an inverse relationship between marine low cloud $L_{50}$ and $\beta$ (not shown) also noted by Wood and Field, [2011], indicating that variations in $\beta$ explain almost one third of the observed geometrical variance in $L_{50}$. Instances of horizontally extensive (greater than 10 km) majority optically thin clouds are observed (Figure 7) but with an order of magnitude smaller frequency than majority optically thick clouds at these lengths. Most notably, poleward of 40° both hemispheres, majority optically thin $L_{50}$ ranges from 5 km to greater than 10 km (Figure 8b). Cloud cover in these regions is mainly Sc (Warren et al., 2010, Climatic atlas of clouds over land and ocean http://www.atmos.washington.edu/CloudMap/WebO/index.html). These relatively large $L_{50}$ values are consistent with the dimensions of clouds in shallow open cellular convection [Wood and Hartmann, 2006]. Majority optically thick and majority optically thin $L_{50}$ maps support our earlier statement that optically thin marine low clouds are in general, smaller in size than optically thick marine low clouds (Figures 8b and 8c).

3.2.2. Cloud Length Distribution From HSRL

[26] The horizontal separation between consecutive CALIOP pulses is 0.335 km but many small cumuli are only a few tens to hundreds of meters in size [Plank, 1969; Hozumi et al., 1982; Wielicki and Welch, 1986; Koren et al., 2008]. The interprofile region not sampled by CALIOP may be populated with one or more small clouds with clear gaps, or may be entirely clear. For our cloud length estimates, if two consecutive profiles are flagged as cloudy we assume that the intervening atmosphere is cloudy, and this could lead to CALIOP overestimating cloud lengths. To examine the extent to which this may occur, we compare spatially matched CALIOP and HSRL cloud length data, noting that HSRL footprints are contiguous. Data are from the tropical western Atlantic trade wind region (section 2.2 for data set description).

[27] A minimum transect length criterion of 30 km is applied to the HSRL-CALIOP comparison data set for estimation of cloud lengths. This is shorter than for our analysis from CALIPSO alone (100 km), because the ±15 min temporal matching limits transect lengths to less than 200 km, and we find that a minimum length requirement of 100 km greatly reduces the data available for comparison in what is already a relatively small data set of collocated data. However, this work (Figure 8) and other studies indicate that characteristic cloud lengths in the tropical western Atlantic region are substantially smaller than 30 km [Benner and Curry, 1998; Wielicki and Welch, 1986; Zhao and Di Girolamo, 2007], and so our reduced transect length threshold for the HSRL-CALIOP size distribution comparison should not produce large problems.

[26] The CALIOP and HSRL cloud length distributions are broadly similar for sizes larger than 1 km, but CALIOP samples more large clouds (Figure 9). This is because HSRL is able to sample smaller clouds than CALIOP can resolve and such clouds comprise 55% of the clouds sampled by HSRL. Subsampling HSRL as CALIOP (i.e., we subsample HSRL profiles every ~335 m) shifts the HSRL cloud length distribution to larger sizes, closely matching the CALIOP data (Figure 9), suggesting that the size distribution difference at cloud lengths smaller than 1 km is primarily due to sparse sampling by CALIOP rather than FOV size differences (HSRL FOV is 60 m; CALIOP FOV is 90 m). CALIOP and HSRL mean $f_{\text{cld}}$ values (0.24 and 0.31, respectively) agree to within 2 standard deviations, but $f_{\text{thin,cld}}$ values do not agree, being 0.89 (CALIOP) and 0.61 (HSRL). This effect cannot be due to the relative along-track sparsity of CALIPSO sampling. Differences in HSRL and CALIOP cloud detection algorithms may be partly responsible for the $f_{\text{thin,cld}}$ disparity (e.g., the HSRL layer detection method may not be well suited to detecting attenuated surface returns). Another likely cause is the narrow HSRL FOV (8 m across track compared with 60 m along track), giving rise to the possibility that HSRL will miss some small clouds that lie on either side of the HSRL FOV, but which lie inside the CALIOP FOV (see schematic in Figure 9 (inset). Since smaller clouds tend to be optically thin, the clouds missed by HSRL are likely to be optically thin.

[29] The HSRL size distribution in Figure 9 shows a roll-off at cloud lengths below about 0.5 km. This may be a hint that there exists a lower limitation to the cloud length of the smallest clouds, i.e., a scale break in the cloud length distribution. We note that this length is consistent with the typical length scale of turbulent eddies in the boundary layer. Without further evidence from observations and models we cannot be sure if the scale break is a real physical feature or reflects sampling a mixture of distributions on different days, but very high-resolution large-eddy simulations could be used to examine this. We note that one high-resolution Landsat satellite study [Sengupta et al., 1990] reported a scale break in cumulus cloud length distribution at 1 km, with larger (smaller) $\beta$ observed for larger (smaller) clouds. The scale break was observed at a larger size (1 km) than in the HSRL data shown here (~500 m). Such a scale break is not a universal feature in trade cumulus [Zhao and Di Girolamo, 2007]. Further studies with high-resolution airborne lidar would be helpful to understand any physical limits to the population of the smallest clouds.

[26] Cloud lengths at median $f_{\text{cld}}$ for the matched HSRL-CALIOP data indicate that $L_{50}$ is 1.3 km for HSRL and 1.7 km for CALIOP. The $L_{50}$ value from 2 years of CALIOP data for this region is 1.3 km, close to the matched data set value. This is also consistent with high-resolution airborne imaging radiometer measurements in other trade wind regions such as the Indian Ocean [McFarquhar et al., 2004]. We therefore believe that the conclusions we derive (see section 5) from the HSRL-CALIOP comparison are somewhat representative of trade cumulus regions throughout the tropics.
3.2.3. Cloud Length Distribution and Optically Thin Cloud Cover Variation

Another approach to examine how optically thin profiles are distributed among clouds of different sizes is to bin, as a function of cloud length \( L \), the fraction of cloudy profiles that are optically thin \( f_\text{thin, cld} \). For this we use the entire CALIOP cloud data set (1.7 million cloud entities).

Surprisingly, the function \( f_\text{thin, cld} \) is almost region invariant (Figure 10). As cloud length increases, the fraction of optically thin profiles comprising that cloud decreases. Clouds of a given horizontal size possess a similar fraction of optically thin elements regardless of whether they occur in regions dominated by trade wind cumulus or stratocumulus. Approximately 50% of profiles comprising clouds of size...
Clouds smaller than 2 km are almost entirely optically thin, whereas clouds larger than 100 km are less than 20% optically thin. A similar relationship is observed in the HSRL data (not shown), but with the HSRL function $f_{\text{thin}, \text{cld}}$ curve shifted to smaller cloud lengths for reasons discussed in the preceding section.

The near region-invariant nature of $f_{\text{thin}, \text{cld}}$ can be explained if each region essentially experiences low clouds of all types, from small cumulus to relatively large sheets of stratocumulus. Visual inspection of visible imagery suggests that this is a reasonable supposition. Surprisingly, each of these cloud types appears to have very similar characteristics in terms of its optical thickness distribution, regardless of location. Clearly, however, the frequency with which each region experiences clouds of different sizes varies dramatically from region to region. This result leads us to ask whether geographical variations in the fraction of clouds that are optically thin might be largely explained by regional differences in the cloud size distribution.

To test this, we make an estimate of the optically thin fraction of marine low clouds ($f_{\text{thin}, \text{cld}}$) using a single $f_{\text{thin}, \text{cld}}$ function based on the entire nonpolar ocean data set (Figure 10) together with the $5^\circ \times 5^\circ$ binned 2 year cloud length distributions $n(L)$ as,

$$f_{\text{thin}, \text{cld}}(L) = \frac{1}{\int_{L_{\text{max}}}^{L_{\text{min}}} n(L)\, dL} \int_{L_{\text{max}}}^{L_{\text{min}}} n(L)\, dL$$

where $L_{\text{max/min}}$ is the upper/lower CALIOP size range limit and $L$ is the bin center cloud length. We find that over nonpolar oceans and separately over the tropics, the cloud length distribution alone explains three quarters of the variance in $f_{\text{thin}, \text{cld}}$ ($R^2 = 0.73$ and 0.77, respectively). What this result tells us is that knowledge of how the marine low-cloud length distribution of all clouds varies is sufficient to accurately predict the geographical variation in the optically thin fraction of clouds across most of the ocean. This is important because it means that the occurrence of optically thin clouds is inherently tied to the size distribution of all low clouds. Further, it suggests that the processes controlling optically thin clouds should not be considered as independent of the processes controlling low clouds in general.

$\sim 15$ km, as determined by CALIOP, are optically thin. Clouds smaller than 2 km are almost entirely optically thin, whereas clouds larger than 100 km are less than 20% optically thin. A similar relationship is observed in the HSRL data (not shown), but with the HSRL $f_{\text{thin}, \text{cld}}$ curve shifted to smaller cloud lengths for reasons discussed in the preceding section.

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3.3. Cloud Vertical Structure

Knowing the vertical distribution of cloud condensate is important for radiative transfer calculations and for model parameterization of cloud overlap. Vertically resolved (60 m) optically thin and optically thick CTH, optically thin cloud base height (CBH), and optically thin cloud layer depth are presented for each of the three regions/regimes discussed previously (section 3.2; Table 3). Note that CBH cannot be observed for optically thick clouds as, by definition, the lidar signal is fully attenuated by the cloud. The dynamical processes associated with each cloud regime are as follows: The vertical development of marine stratocumulus is suppressed by subsidence associated with either the descending branch of the Hadley/Walker circulation or subsiding air masses associated with midlatitude cyclones, depending upon location. As a result, the marine stratocumulus layer is capped by a strong temperature inversion, resulting in extensive low clouds [Klein and Hartmann, 1993]. As the stratocumulus air mass moves equatorward it is advected over warmer sea surface temperatures (SSTs), surface latent heat fluxes increase, and the subsequent transition to a cumulus regime is explained by the deepening-warming mechanism proposed by Wyant et al. [1997]. Consistent with a deepening marine boundary layer from stratocumulus to trade cumulus, both optically thin and optically thick CTH distributions peak at higher altitudes transitioning from a stratiform to a convective type cloud regime (Figures 11a–11c).

In the stratocumulus region, optically thick CTH peaks at twice the altitude observed for optically thin clouds (1.2 km and 0.6 km, respectively). Although cloud cover in stratocumulus regions is dominated by optically thick cloudy profiles (Figure 5c), optically thin cloudy profiles comprise greater than 95% of clouds up to 4 km in size (Figure 10). Together, this suggests that small clouds typically have lower CTH than larger, optically thicker clouds. Noting that estimates of stratocumulus depth range from 200 to 300 m in the subtropics and tropics [Wood, 2011], and that optically thin cloud layers preferentially reside in the lowest 1 km with mean depth of 200 m (Figure 12a), we suggest that incidences of optically thin CTH below 1 km in this stratocumulus region are separate cloud entities from the optically thicker clouds with CTH greater than 1 km. Many of these likely reflect expansive collapsed boundary layers and fog layers (note occurrence of very low cloud bases in Figure 11d and data along the 1:1 line in Figure 12a) that occur periodically in regions dominated by marine stratocumulus [Ackerman et al., 1993; Christensen and Stephens, 2011].

As would be expected given the weakening of the capping inversion and deepening of the marine layer in the...
transition from stratocumulus to cumulus, optically thin and optically thick CTH distributions are shifted upward relative to the stratocumulus regime (see Figures 11a and 11b). Cloud cover remains largely optically thick, but in contrast to the stratocumulus region the tops of these layers tend to occur at altitudes closer to the optically thick layers, suggesting that the optically thin clouds in the transition region may be largely associated with thinning stratocumulus and stratus layers and remnants of dissipating stratocumuli near the top of the boundary layer.

In the trade wind cumulus region (Figures 11c and 11f), the modal heights for optically thin and optically thick cloudy profiles are again at different levels (0.8 and 2 km respectively). Optically thin profiles dominate the cloud cover in this regime ($f_{\text{thin,cld}} = 0.84$), and clouds are not horizontally extensive ($L_{50} = 0.96$ km). Despite having modes at distinct levels, optically thin and optically thick CTH are observed above 400 m up to a height of 3 km (Figure 11c). Peak optically thin CBH is ~600 m (Figure 11f). Detrained elements from optically thick clouds may account for a large proportion of optically thin clouds at altitudes greater than 1 km, but the mode in optically thick clouds below 1 km probably consists of small cumuli and associated fragments of clouds with insufficient buoyancy to grow in stature. An aerial photograph of a trade wind cloud field (Figure 13) highlights the wide range of horizontal and vertical cloud dimensions found in trade wind regions. Notable is the prevalence of tenuous, wispy clouds with limited vertical extent, along with optically thicker clouds, although the latter are less numerous. The bimodal nature of the combined optically thin and optically thick cumulus CTH distributions (sum of both distributions, not shown) agrees well with high-resolution (90 m) satellite

![Figure 12. Normalized density distributions of optically thin cloud top height and layer depth for three cloud types (region center latitude, longitude): (a) California stratocumulus region (25°N, 125°W), (b) tropical Pacific stratocumulus to cumulus transition (Sc-Cu Trans; 15°N, 135°W), and (c) tropical Pacific trade wind cumulus (Cu; 15°S, 155°W). The insets are layer depth distributions for each region, overlaid with a gamma distribution (red line). The values are the region mean (standard deviation) layer depth (km) and number of full-resolution optically thin cloud profiles for each region. The layer depth histogram is normalized such that the area under the curve equals 1.](image-url)
observations of trade wind cumulus CTH [Genkova et al., 2007] and with a large-eddy simulation (LES) cloud cover distribution for a nonprecipitating trade cumulus case [Siebesma et al., 2003]. This encouragingly suggests that LES is a useful tool to examine macrophysical properties of optically thin clouds that are the dominant cloud type in the trade wind cloud regimes.

[38] The optically thin cloud layer thickness distributions (Figure 12, insets) are similar for all three regions. The drop-off in frequency of occurrence of layers less than 180 m may be an artifact of the increase in the nighttime optical depth detection threshold for shallow layers, i.e., the threshold is 0.03 for layers greater than 180 m, and 0.1 for shallower layers (section 2.1). We note, however, that optically thin cloud layer depth may be overestimated by CALIOP VFM: if a cloud partially fills a 60 m height bin yet has an optical depth greater than 0.1, the entire 60 m bin will be classified as cloudy. This high bias will affect shallow layers more than deep layers, in a relative sense. For example, a 60 m optically thin cloud partially filling two vertically adjacent 60 m bins, will register as 120 m in depth, an error of 100%, whereas, an optically thin cloud layer occupying four 60 m bins, and partially filling a fifth bin by 50% will have a layer depth relative error of ~11%. Conservatively assuming the maximum error of 60 m in layer depth yields a 25% potential overestimate of mean optically thin cloud layer depth. Irregular or inhomogeneous clouds e.g., trade cumulus, will be most prone to this error. A recent LES study investigating the effects of cumulus cloud irregularity on cloud overlap quantified the effect of partially filled model levels as a function of model vertical resolution [Neggers et al., 2011]. Employing their formulation at CALIOP VFM cloud layer boundary detection resolution (60 m) produces a high-bias estimate of 28%, which is very similar to our own estimate.

4. Discussion

[39] Our findings point to a prevalence of optically thin low clouds over the oceans, even at the scale of individual CALIOP lidar profiles. While this might be expected at the scale of typical moderate resolution passive visible remote sensors used to construct cloud climatologies, our prior expectation was that most of the optically thin cloudy pixels from these moderate-resolution passive sensors would be partially filled by small cumulus covering only a small fraction of the pixel, and that the clouds themselves might have considerably higher optical thicknesses than the pixel as a whole. This appears not to be the case, and it prompts us to ask whether even the CALIOP FOVs are partially filled with yet smaller clouds. Given that there are many single-profile cloud entities in the CALIOP data set (Figures 7 and 9), it is possible a significant fraction of these profiles are partially filled.

[40] The cloud within any CALIOP partially filled field of view (pFOV) could be optically thick or optically thin. Based upon CALIOP data alone, we cannot diagnose whether or not a cloudy FOV is partially cloud filled, nor can we determine if the cloud-filled part is optically thick or optically thin. Classifying a pFOV as optically thin when the cloud intercepted is optically thick cloud will lead to an overestimate of  $f_{\text{thin,cld}}$. To investigate the frequency of pFOVs, and whether or not the clouds partially filling the lidar FOVs are optically thick, we employ a simple bounded cascade fractal model.

4.1. Bounded Cascade Model Description

[41] The theoretical basis for, and an in-depth description of the model is presented by Marshak et al. [1994] and its application to simulating one-dimensional cloud fields is provided by Wood and Field [2011]. A brief description follows. A one-dimensional multiplicative cascade broken cloud field is generated by initially assuming a uniform value of a cloud property; here, optical depth. Irregular or inhomogeneous clouds e.g., trade cumulus, will be most prone to this error. A recent LES study investigating the effects of cumulus cloud irregularity on cloud overlap quantified the effect of partially filled model levels as a function of model vertical resolution [Neggers et al., 2011]. Employing their formulation at CALIOP VFM cloud layer boundary detection resolution (60 m) produces a high-bias estimate of 28%, which is very similar to our own estimate.

\[
\text{Figure 13. An aerial photograph of a typical trade wind cumulus cloud field observed in the western Caribbean during the Rain in Cumulus over Ocean (RICO) field program. The wide range of cloud depths and horizontal extents is apparent. Image courtesy of B. Stevens.}
\]
model has displayed skill in representing scaling properties of marine stratocumulus liquid water content [Marshak et al., 1997; Davies et al., 1999], and regional/global cloud size statistics derived from the cloud optical depth field [Wood and Field, 2011]. Specifically, it is shown in the latter case that the power law exponent $\beta$ in the cloud chord length distribution is inversely related to the value of $H$ controlling the optical depth field, such that smooth fields (large $H$) are associated with more large clouds and low $\beta$, and vice versa.

The model is used to generate 100 optical depth realizations, each with 17 cascade steps and therefore 218 individual data points of horizontal scale 2.8 m and total length 700 km. This large range of scales is chosen so that we can (a) use the model to explore sub-FOV variability within the CALIOP FOV (we have 32 model pixels in the 90 m FOV) and (b) have sufficiently long transects to provide good statistics and be approximately consistent with the aggregation scale for our CALIOP analysis. Consecutive FOVs are separated by 335 m to be consistent with CALIOP sampling. We correct all modeled cloud length distributions for sampling bias associated with clouds approaching the limited transect length using equation (1).

The optical depth threshold for CALIOP cloud detection ($\tau_{\text{cld,thresh}} = 0.03$) is applied to determine if a given model FOV is cloudy. If a surface is detected by CALIOP in the presence of a cloud, it is not known a priori whether the cloud is optically thin, or if an optically thick cloud is partially filling the FOV. To determine if CALIOP would detect the surface for any given cloudy model FOV (i.e., to determine if the FOV would be classified as optically thin cloud), we compare the number of surface-returned photons ($n_s$) that would be obtained from a model-simulated cloudy FOV to the minimum number of surface-returned photons ($n_{\text{rs}}$) required to detect the surface. If the surface returns $N$ photons without cloud attenuation, then from Beer’s law, $n = N e^{-\tau_s}$. This ignores multiple scattering, which is weak for optically thin clouds. We know that the surface is just detected in the presence of a uniform cloud with $\tau_s = 3$, and so $n_s = N e^{-2\tau_s}$, which gives $n/n_s = e^{-2(\tau_s-\tau)}$. We compute $n/n_s$ for every model pixel (2.8 m scale) and then take the mean over the CALIOP FOV. If this FOV mean $N/N_s > 1$ then the surface would be detected and such a cloudy FOV is classified as optically thin to best match our CALIOP observational definition.

We focus our investigation on our representative tropical Pacific trade wind region (Table 3) because the trades have the smallest marine low clouds (Figure 8a) and are therefore the most prone to the partially filled FOV problem [Jones et al., 2012]. Experimentation shows that model settings best simulating CALIOP observations are $p = 0.0005$, $H = 0.07$, with scene mean optical depth (also a specified input variable) set to 4.5, selected as follows. Because $H$ is closely related to $\beta$ (see Figure 15) [Wood and Field, 2011], we vary $H$ to match CALIOP observed $\beta$ (Figure 14). We also vary $p$ to match observed CALIOP trade wind region mean $f_{\text{cld}}$ and $f_{\text{hstd,clrd}}$ as closely as possible. Finally, scene mean optical depth is adjusted such that TOA albedo $\alpha_{\text{TOA}}$ estimated from model $\tau_{\text{cld}}$ values, falls within the range of CERES albedo values for this region. Calculation details are presented in full in Table 4.

### 4.1.1. Model Simulation of Observed Cloud Properties

The model demonstrates skill in generating realistic representations of CALIOP observed cloud properties $f_{\text{cld}}$ and $f_{\text{hstd,clrd}}$, $\beta$ and TOA albedo $\alpha_{\text{TOA}}$ (Table 4). Consistency between observed (CALIOP) and predicted cloud size distributions suggests that optically thin cloud behavior is broadly consistent with a power law scaling of optical depth: this scaling is an underlying assumption of the model. For comparison, we also sample the model output simulating MODIS and HSRL FOV (1 km and 60 m along-track dimensions, respectively). HSRL-simulated model output is used the same $\tau_{\text{cld,thresh}} = 0.03$ and $n/n_s > 1$ thresholds used for CALIOP. We set MODIS $\tau_{\text{cld,thresh}}$ to be 0.4 [Ackerman et al., 2008], with $\tau_{\text{hstd,thresh}}$ as for CALIOP. Perhaps unexpectedly, model-simulated MODIS $f_{\text{cld}}$ is 1.5 times lower than CALIOP- and HSRL-simulated $f_{\text{cld}}$ (Table 4) Varying MODIS $\tau_{\text{cld,thresh}}$ within reasonable limits (0.2 to 0.6) produces model simulated MODIS $f_{\text{cld}}$ values of 0.20 and 0.14 respectively (i.e., still lower than the CALIOP value of 0.27). We therefore suggest that overestimation of cloud cover using passive visible imagers [e.g., Zhao and Di Girolamo, 2006] may not necessarily always occur. Rather, the bias depends sensitively upon cloud length and the reflectance threshold applied.

As instrument resolution increases (FOV size decreases), each simulated size distribution is shifted to smaller sizes (Figure 14). We note that the model-simulated HSRL cloud length distribution, since it is derived from a bounded
where the number of photos required for surface detection is greater than one, the cloudy FOV is classified as optically thin. The mean n/n_{\text{cld}} for the entire pFOV is calculated, along with the mean optical depth of the cloud-only segment. For a given pFOV with n/n_{\text{cld}} > 1 and mean optical depth of the cloud-only segment greater than 3, the pFOV is therefore misclassified as an optically thin cloud whereas it is really a partially filled FOV containing an optically thick cloud. The frequency of such misclassification is estimated. Results are binned by cloud length, and presented in Figure 15.

The fraction of FOVs that are partially cloud filled is high, especially for smaller clouds where it exceeds 0.8. This

4.2. Partially Filled Fields of View

As previously mentioned, a CALIOP FOV may be misidentified as optically thin if partially filled by an optically thick cloud. The model is employed to identify simulated CALIOP cloudy FOVs that are partially cloud filled (pFOV), i.e., not all 32 optical depth values comprising the FOV exceed the threshold for cloud detection (\(\tau_{\text{cld,thresh}}\)). As discussed in section 4.1, if the FOV mean ratio of surface-backscattered photons detected by the sensor (n) to the minimum number required for surface detection (n_{cld}) is greater than one, the cloudy FOV is classified as optically thin. The mean n/n_{\text{cld}} for the entire pFOV is calculated, along with the mean optical depth of the cloud-only segment. For a given pFOV with n/n_{\text{cld}} > 1 and mean optical depth of the cloud-only segment greater than 3, the pFOV is therefore misclassified as an optically thin cloud whereas it is really a partially filled FOV containing an optically thick cloud. The frequency of such misclassification is estimated. Results are binned by cloud length, and presented in Figure 15.
fraction decreases to around 0.3 for larger clouds (Figure 15, circles). CALIOP-observed cloud entities consisting of a single profile are therefore almost all likely to be broken clouds. The bin median pFOV cloud fraction, i.e., fraction of a pFOV that is cloud, increases with cloud length, ranging from 0.50 to 0.85 (Figure 15, squares), consistent with fewer cloud edges being observed as cloud length increases. Despite the prevalence of partially filled FOVs, the FOVs classified as optically thin typically do contain mainly optically thin cloud (Figure 15, diamonds). Summing the fraction of pFOV that are misclassified for each size bin, and weighting this value by the fraction of optically thin FOVs that are partially filled, it can be concluded that the potential overestimate of \( f_{\text{thin,cld}} \) is at most 6% for this cloud regime. In addition, the effect of sensor resolution on \( f_{\text{cld}} \) can be quantified from the model output. The overestimate in \( f_{\text{cld}} \) caused by assuming that partially cloud-filled FOVs are 100% cloud filled is estimated at \( \sim 0.06 \) in absolute terms. This value is similar to the ratio of model generated CALIOP and HSRL Full Res \( f_{\text{cld}} \) (0.27/0.22 = 1.23), with the latter representing “true” \( f_{\text{cld}} \).

[50] Based upon the model output, which is designed to simulate typical trade wind cumulus conditions, it is concluded that the majority of cloudy CALIOP profiles are indeed partially cloud filled. Rapid fluctuations of cloud particle concentration on scales \( \sim 10 \) m have been measured in situ [Pawlowska et al., 1997], suggesting that this conclusion is physically realistic. These results have implications for the minimum sensor resolution required for cloud observations in regions of broken cloud, as simulated by the model. A recent study suggests that a resolution higher than 45 m is required to accurately determine the cloud cover to within 0.01 for trade cumulus regimes [Jones et al., 2012]. Our study suggests that the CALIOP FOV size of 90 m leads to an overestimate of some 0.06 in cloud cover in regions of trade cumulus. This is consistent with Jones et al. [2012], whose results (their Figure 1) suggest a median overestimation of 0.05 for a sensor with a resolution of \( \sim 100 \) m.

4.3. Additional Applications of Cascade Model Output

4.3.1. Are Optically Thin Clouds Adiabatic?

[51] We analyze the model output to investigate the extent to which optically thin clouds might be considered adiabatic, i.e., liquid water path (LWP) values determined by temperature, pressure and cloud thickness alone. Liquid water path for an adiabatic cloud is proportional to cloud layer depth (\( h \)) squared, the adiabatic rate of increase of liquid water with respect to height (assumed to be \( 2 \times 10^{-3} \) g m\(^{-3}\) km\(^{-1}\) [Albrecht et al., 1990], and the adiabatic fraction \( f_{\text{cld}} \), is defined as the ratio of observed LWP to adiabatic LWP, which here we assume to be 1. Using model-simulated optically thin cloud optical depth for a trade wind region as before, and assuming a cloud droplet effective radius \( r_e \) of 10 \( \mu \)m, we estimate LWP = \( 2/3 \pi \rho_L r_e \) [Stephens, 1994], where \( \rho_L \) is liquid water density, and derive corresponding adiabatic cloud layer depth, \( h \). The simulated adiabatic layer depth distribution (not shown) indicates that depths are on the order of a few tens of meters, with 90% of the values being less than 90 m, much thinner than observed. In addition, if this LWP is evenly distributed over a 60 m layer, we estimate that \( f_{\text{cld}} \sim 0.05 \), and ranges from 0.02 to 0.09 when \( r_e \) is halved and doubled, respectively. Alternatively, if \( \tau_{\text{cld}} \) is 0.1, layer depth corresponding to an adiabatic LWP is 25 m.

Therefore, we conclude that the optically thin cloud layers are not adiabatic, but are more likely to be diluted cloud fragments, and their LWP should not be estimated adiabatically. This is consistent with effects of entrainment on marine layer clouds [e.g., McFarlane and Grabowski, 2007]. Stochastically controlled entrainment events essentially dictate the properties of shallow trade cumulus clouds [Romps and Kuang, 2010].

4.3.2. Optically Thin Cloud Contribution to TOA Cloud Albedo

[52] We have shown that optically thin clouds comprise almost one half of all marine low clouds over nonpolar oceans. To investigate the magnitude of their contribution to low-cloud \( \alpha_{\text{TOA}} \) in trade wind regions at CALIOP FOV resolution, model-simulated optical depth is converted to albedo using a simple approximation, \( \alpha_{\text{TOA}} = \tau_{\text{cld}} (\tau_{\text{cld}} + 7.7) \) [Lacis and Hansen, 1974], and \( \alpha_{\text{TOA}} \) is estimated for optically thin clouds and all low clouds (Table 4). The ratio of these values is expressed per unit of low-cloud cover and per unit of optically thin cloud cover, and the calculation is repeated at model full-resolution to investigate how optically thick clouds classified as optically thin clouds in pFOVs, affect the results.

[53] In addition to the marine cumulus region described in Table 3, two other trade cumulus regions (of equal size) were examined; one in the tropical western Pacific and one in the tropical western Atlantic. For trade wind regions it can be concluded that with observed \( f_{\text{thin,cld}} > 0.90 \), and \( f_{\text{cld}} \sim 0.20–0.30 \), optically thin low clouds can contribute 55–60% to the total low-cloud albedo. However, the range of values obtained from the full-resolution model simulation indicates that the true contribution of optically thin clouds to low-cloud albedo is more than a factor of two smaller at 23%–26%.

5. Summary and Concluding Remarks

[54] Analysis of 2 years of CALIOP nighttime data indicates that 45% of marine low clouds with tops below 3 km, over the nonpolar oceans, are classified as optically thin such that they do not fully attenuate the CALIOP lidar signal. Few observational data, other than this work, are available on the physical nature of these clouds. Key results from our analysis are as follows.

[55] 1. Over the nonpolar oceans, optically thin clouds comprise 0.25–0.75 of marine low clouds with cloud top height below 3 km, with a mean of 0.45.

[56] 2. The optically thin fraction of marine low cloud varies inversely with marine low-cloud cover, and reaches a maximum (>0.80) in trade wind regions.

[57] 3. Although the optically thin fraction of low clouds peaks in trade wind regions, the absolute frequency of occurrence does not strongly vary over the nonpolar oceans.

[58] 4. In trade wind regions, according to our analysis of CALIOP observations, clouds smaller than 2 km contribute over 50% of the cloud cover.

[59] 5. Optically thin marine low clouds are predominantly small clouds. The cloud length distribution of all clouds explains three quarters of the geographical variance in the optically thin fraction of marine low clouds.

[60] 6. Over the nonpolar oceans, optically thin low clouds have a mean cloud top height of 1.2 km and a mean thickness of 0.25 km. However, optically thin cloud layer depth
may be overestimated by 25% due to cloud partially filling the 60 m vertical resolution CALIOP bins.

[61] 7. Collocated HSRL and CALIOP data for the tropical western Atlantic trade wind region reveal that many clouds are smaller than can be resolved at CALIOP resolution. This suggests that clouds, especially optically thin low clouds, may exist globally on finer scales than reported herein.

[62] 8. A simple bounded cascade fractal model is used to represent a one-dimensional optical depth distribution for conditions representative of the trade wind regions. Analysis of model output, in conjunction with observations, suggests that (a) optically thin low-cloud properties are consistent with a power law scaling of optical depth, (b) CALIOP fields of view partially filled with optically thick low clouds produce a potential 6% (relative) overestimate in optically thin fraction of marine low-cloud values in trade cumulus regions, (c) optically thin cloud layer thicknesses are not adiabatic with greater than 90% of simulated optically thin adiabatic cloud depths less than 90 m, whereas observations indicate much greater thicknesses, and (d) in trade cumulus regions, we estimate that optically thin low clouds contribute almost a quarter of the albedo contributed by low clouds, but a more sophisticated quantification using radiative transfer modeling is required to confirm this.

[63] Remote sensing of clouds provides global information on cloud macrophysical and microphysical properties, both of which are important for radiative transfer calculations and verification of model representation of clouds. Differences in cloud properties retrieved by ISCCP and MODIS [Pincus et al., 2012] and ISCCP, MODIS and MISR [Marchand et al., 2010] highlight the difficulties inherent in passive remote sensing of low clouds. Pincus et al. [2012] reported that ISCCP and MODIS views of clouds are consistent for all but the optically thinnest clouds, while subpixel scale and broken low-level clouds were identified by Marchand et al. [2010] as one of the reasons for discrepancies in cloud properties.

[64] It is known that low clouds are the dominant contributor to the global albedo [Hartmann et al., 1992]. Results from our work indicate that almost half of low clouds over the nonpolar ocean are optically thin. Additionally, in trade wind regions, 50% of the low-cloud cover is composed of clouds less than 2 km in length. Because such a large fraction of cloud is optically thin, even with CALIOP’s 90 m footprint size, it is doubtful that the passive sensors are actually determining the coverage of the optically thick or optically thin clouds correctly. It is difficult therefore, to conclude that we yet have a precise understanding of the shape of the true cloud optical depth probability distribution function on a global scale, and especially in trade wind regions.

[65] Although CALIOP data do not provide vertically resolved high-resolution off-track spatial context for the cloud fields sampled, such data in the along-track direction afford new insights on the nature of optically thin clouds. Suggestions for design of future combined active and passive remote sensors would include an array of lasers providing a swath of vertically resolved data with, ideally, horizontal resolution of a few tens of meters or less, in the across-track and along-track directions, and collocated visible imagery with the same horizontal resolution. Such a configuration would provide collocated profiles of cloud properties, and high-resolution TOA reflectance data to better observe these optically thin clouds, that comprise almost half of all marine low clouds.

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