Ultraclean Layers and Optically Thin Clouds in the Stratocumulus-to-Cumulus Transition. Part I: Observations

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ABSTRACT

A common feature of the stratocumulus-to-cumulus transition (SCT) is the presence of layers in which the concentration of particles larger than 0.1 \( \mu \)m is below 10 cm\(^{-3}\). These ultraclean layers (UCLs) are explored using aircraft observations from 14 flights of the NSF–NCAR Gulfstream V (G-V) aircraft between California and Hawaii. UCLs are commonly located in the upper part of decoupled boundary layers, with coverage increasing from less than 5% within 500 km of the California coast to \( \sim 30\%\)–60% west of 130°W. Most clouds in UCLs are thin, horizontally extensive layers containing drops with median volume radii ranging from 15 to 30 \( \mu \)m. Many UCL clouds are optically thin and do not fully attenuate the G-V lidar and yet are frequently detected with a 94-GHz radar with a sensitivity of around \( \sim 25\) dB\(Z\). Satellite data indicate that UCL clouds have visible reflectances of \( \sim 0.1\)–0.2 and are often quasi laminar, giving them a veil-like appearance. These optically thin veil clouds exist for 1–3 h or more, are associated with mesoscale cumulus clusters, and likely grow by spreading under strong inversions. Active updrafts in cumulus (Cu) clouds have droplet concentrations of \( \sim 25\)–50 cm\(^{-3}\). Collision–coalescence in the Cu and later sedimentation in the thinner UCL clouds are likely the key processes that remove droplets in UCL clouds. UCLs are relatively quiescent, and a lack of mixing with dry air above and below the cloud may help to explain their longevity. The very low and highly variable droplet concentrations in UCL clouds, together with their low geometrical and optical thickness, make these clouds particularly challenging to represent in large-scale models.

1. Introduction

Assessment of the physical factors controlling the coverage and albedo of marine boundary layer (MBL) clouds remains a pressing challenge. The speed at which the stratocumulus (Sc)-to-cumulus (Cu) transition (SCT) occurs in air masses downstream of the eastern subtropical ocean basins determines the albedo of the tropics, and similar cloud transitions in postfrontal air masses are important for determining midlatitude stormtrack albedo. These transitions in cloudiness are poorly represented in climate models, which typically show stratocumulus breakup occurring too quickly in both the subtropics (Teixeira et al. 2011) and the mid-latitudes (Williams et al. 2013; Bodas-Salcedo et al. 2014). In addition to producing too few low clouds, model clouds are too reflective, a pathology commonly referred to as the “too few, too bright” problem (Nam et al. 2012; Engström et al. 2015).

In general, the SCT initiates in response to MBL deepening and decoupling as surface fluxes increase over warmer waters (Bretherton and Wyant 1997; Wyant et al. 1997). As several studies have shown (Martin et al. 1995; Zhou et al. 2015), stratocumulus cloud breakup is not an immediate response to MBL decoupling but can be delayed by as much as 1–3 days with cloud cover often remaining above 50% until 500–2000 km downstream of decoupling onset (Zhou et al. 2015). Given the

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importance of the SCT, it is remarkable that few dedicated aircraft observations have sampled the MBL in the breakup region owing to the typically remote location of this phenomenon and the limited range of most research aircraft. Early transition experiments with shipborne observations [e.g., the 1969 Airmass Transition Experiment (ATEX; Augstein et al. 1973; Stevens et al. 2001)] provided important characterization of the vertical thermodynamic structure of the MBL under transition conditions but little if any information on cloud structure and dynamics. Aircraft observations made during the Atlantic Stratocumulus Transition Experiment (ASTEX) in 1992 provide one of the most widely used datasets for understanding the SCT and are perhaps the first measurements of cloud microphysical properties in the SCT. The two Lagrangian transition experiments carried out in ASTEX showed that drizzle occurred with sufficient frequency and rate to be important in MBL energy and moisture budgets (Bretherton and Pincus 1995; Bretherton et al. 1995). Surface-based radar measurements in ASTEX also highlighted the frequent occurrence of precipitating cumulus clusters with horizontal scales of \( \sim 10 \text{ km} \) rising into and interacting with stratocumulus clouds at the top of the decoupled MBL (Miller and Albrecht 1995). Martin et al. (1995) showed that the stratocumulus deck is locally thickened in regions of interaction with rising Cu, consistent with the Cu providing a source of moisture to the decoupled cloud layer aloft. However, the Cu also drive entrainment of dry air from the free troposphere, which eventually serves to dry the upper MBL and dissipate the Sc (Krueger et al. 1995; Wyant et al. 1997). The time scale over which drying happens (i.e., the time between decoupling and stratocumulus breakup) likely depends upon the relative amount of free-tropospheric entrainment that is achieved for a given Cu mass flux, the inversion strength (which influences both entrainment and the spreading of Cu below the inversion), the free-tropospheric humidity, and the rate at which the upper MBL loses condensate through precipitation and droplet sedimentation. The processes causing drying therefore are expected to depend upon the nature of the clouds and turbulent dynamics in the upper MBL.

Large-eddy simulations (LESs) indicate that the SCT is sensitive to cloud microphysical processes in addition to large-scale meteorological controls. Sandu and Stevens (2011) found that reducing the cloud droplet concentration \( N_d \) (assumed to be constant throughout the domain and over the simulation time) from 100 to \( 33 \text{ cm}^{-3} \) increases precipitation efficiency and slows down the growth of the MBL but hastens the rate of cloudiness reduction. Although not specifically transition experiments, previous short-duration LES studies have shown a strong dependence of stratocumulus cloud cover on \( N_d \) (Ackerman et al. 2003), especially when \( N_d \) falls below \( 50 \text{ cm}^{-3} \). Recently, Yamaguchi et al. (2017) find that \( N_d \) constitutes a strong control on the SCT because of its role in modulating the drizzle process. These results are broadly consistent with the Sandu and Stevens (2011) study, pointing to the need to understand what controls \( N_d \) in the SCT.

Several cases of spatial transitions from closed- to open-cellular convection over the southeastern Pacific Ocean were observed during the VAMOS Ocean–Cloud–Atmosphere–Land Study (VOCALS) Regional Experiment (Terai et al. 2014) and showed that within regions of open cells, active Cu clouds that draw aerosol from the surface mixed layer have \( N_d \) values approximately 3 times larger than the layer of thin stratus/stratocumulus at the top of the MBL into which the Cu are detraining. Further, \( N_d \) in the stratus layer of the open-cell region was found to be extremely low (\( 5 \text{ cm}^{-3} \) in the mean, over five cases), prompting its characterization as an ultraclean layer (UCL), a term first used for clean, thin unpolluted layers between aerosol-rich polluted layers over southern Africa (Hobbs 2003). In one open-cell case during VOCALS, \( N_d \) in the most active Cu updrafts was over an order of magnitude greater than that in the relatively quiescent detrained stratus layer (Wood et al. 2011). Other observational studies have found cases with extremely low \( N_d \) and cloud condensation nuclei concentrations in the subtropical MBL (Stevens et al. 2005; Sharon et al. 2006; Petters et al. 2006; Wood et al. 2008), in midlatitude marine cold-air outbreaks (Field et al. 2014), and in the Arctic (Mauritsen et al. 2011). Combining data from satellite with ship cruises between California and Hawaii, Painemal et al. (2015) showed that the correlation of satellite \( N_d \) and surface-based accumulation aerosol concentration was 0.9 east of 145\textdegree W but dropped to 0.5 in the SCT west of this, consistent with a decoupling of cloud microphysics from near-surface aerosol and surface aerosol sources in the SCT. This would be consistent with an aerosol sink in the upper PBL that is uncorrelated with near-surface aerosol.

The dramatic variability of \( N_d \) even within the same cloud system raises the question of whether fixing \( N_d \) (or aerosol concentration) across an entire model domain is appropriate for the study of the SCT. Recent LES studies (e.g., Kazil et al. 2011; Berner et al. 2013) have allowed \( N_d \) and aerosols to vary spatiotemporally in response to coalescence scavenging and Brownian scavenging, surface particle fluxes, entrainment, and advection. These studies have been able to reproduce with some fidelity the vertical and horizontal variability of \( N_d \) and aerosol concentration in the MBL seen in the
open-cell cases observed in VOCALS. A dramatic reduction in $N_d$ due primarily to coalescence scavenging appears to be a universal feature of the transition from closed to open cells, although the level of sophistication in the aerosol budget necessary to faithfully reproduce the cloudiness transition is unclear. What is also unclear is the extent to which closed-to-open-cell transitions and specifically the formation of UCLs are representative of the behavior of SCTs observed elsewhere.

This study explores the nature of clouds and aerosols in the upper MBL occurring in the SCT, with specific focus on understanding the relationship between UCLs and frequently occurring optically thin clouds in these layers. The primary tool is aircraft observations from 14 flights of the NSF–NCAR Gulfstream-V (G-V) aircraft during the Cloud System Evolution in the Trades (CSET) field program. CSET sampled the MBL and lower free troposphere extending across the entire region from the California coast (~38°N, 123°W) to Kona, Hawaii (19°N, 156°W), a distance of 3700 km. Trajectory calculations indicate that air masses take approximately 5 days to advect this distance (Sandu et al. 2010). Thus, CSET provides unprecedented observations of the complete transition from shallow marine stratocumulus and stratus, to deep well-mixed and then decoupled Sc, through to a largely trade Cu environment at Hawaii, albeit one with 40% coverage by optically thin upper-PBL stratiform clouds.

The remainder of this paper is as follows: Section 2 describes the observational datasets used and the methodological approach. Section 3 examines the statistical occurrence of both clear and cloudy UCLs including their frequency, altitude, and geographical distribution. Section 4 documents properties of both cloudy and clear UCLs using in situ G-V data. Section 5 explores several case studies from CSET, combining in situ and remote sensing data from the G-V in conjunction with satellite remote sensing, to provide further insights into the microphysical and macrophysical structure of UCLs and optically thin clouds. Section 6 documents the statistical properties of optically thin MBL clouds observed in CSET, relating their vertical structure to that of cloudy UCLs. Finally, section 7 introduces hypotheses for how UCLs are formed and discusses some of the implications of the results presented in this study.

2. Data and methods

a. Instruments and sampling

Data from the CSET field campaign (Albrecht 2018, manuscript submitted to Bull. Amer. Meteor. Soc.) constitute the primary observational dataset used in this study. The NSF–NCAR G-V High-Performance Instrumented Airborne Platform for Environmental Research (HIAPER) aircraft sampled marine air masses between Sacramento, California (38.6°N, 121.5°W), and Kona (19.6°N, 156.0°W). The G-V was instrumented with numerous instruments, which are detailed in Albrecht (2018, manuscript submitted to Bull. Amer. Meteor. Soc.) and are only described briefly here. In addition to measurements of the atmospheric state variables (temperature, pressure, water vapor, and winds including high-frequency turbulent components at 20 Hz), other instruments used in this study include the following:

- An ultra-high-sensitivity aerosol spectrometer (UHSAS) measures aerosol particles with diameters from 60 to 1000 nm, with 100 size bins. Here, we use only the most reliable size bins from 100 to 1000 nm and define $N_d$ as the concentration of particles larger than 100 nm. UHSAS data are used only in cloud-free conditions.
- A cloud droplet probe measures the cloud droplet size distribution from 1 to 25-μm radius (with a 1-μm bin width), and a fast two-dimensional cloud optical array probe (2DC) provides a size distribution for larger drops between 37.5- and 800-μm radius with a bin width of 12.5-μm radius. The concentration of drops measured by the cloud droplet probe (CDP) is termed the droplet concentration $N_d$. The CDP and 2DC data are combined to produce a single cloud–drizzle size distribution from 1 to 800 μm, with spline interpolation in log–log space used to fill the gap region between the upper size bin of the CDP and the smallest useable bin of the 2DC. In addition to the CDP and 2DC, the G-V also flew a holographic cloud droplet detector (HOLD-C; Fugal and Shaw 2009) that uses holographic reconstruction to measure droplets spanning the poorly measured gap region.
- A King hotwire probe measures liquid water content.
- A VCSEL water vapor hygrometer and a Rosemount temperature sensor.
- The HIAPER Cloud Radar (HCR), a 94-GHz Doppler radar, can sample either in the nadir or zenith directions (Vivekanandan et al. 2015). Nadir sampling is used when the aircraft is above cloud or in cloud layers, and zenith sampling is used when the aircraft is below the cloud level.
- High-spectral-resolution lidar (HSRL), a 532-nm lidar (Grund and Eloranta 1991), can sample either in near-nadir or near-zenith directions. The lidar is manually turned from nadir to zenith sampling, and the pointing direction is set to be almost the same as that for the HCR except that the HSRL is designed to be pointed 4° laterally off-nadir/zenith to avoid specular reflections from ice crystals.
We use all available data from 14 of the research flights (RFs; RF02–RF15) that were designed to sample air masses transitioning from Sc to Cu. The outbound flights (from Sacramento to Kona) comprised a transit leg (typically at 6–8-km altitude) from Sacramento to an offshore location containing Sc clouds relatively close to the California coast. The G-V then profiled down to conduct low-level sampling of the PBL and lower free troposphere (FT) for a horizontal distance of typically 1500–2000 km before ascending again to ~6 km for the transit to Kona. Low-level sampling constituted approximately half of the flight. The return flight from Kona to Sacramento was conducted 2 days after the outbound flight and was designed to sample the locations of the Lagrangian-advected PBL air masses that had moved toward Hawaii over the 2 days between the outbound and return flights. The return flight consisted of approximately the same amount of low-level sampling as was carried out on the outbound flight. The locations of the flights are shown in Albrecht (2017, manuscript submitted to Bull. Amer. Meteor. Soc.). The low-level-sampling strategy consisted of repeated flight modules comprising a 10-min near-surface straight and level leg (~140 m above the surface), a 10-min straight and level leg 100 m above the cloud base (either the Sc cloud base for overcast conditions or the Cu cloud base for broken cloud conditions), a sawtooth leg porpoising from approximately 100 m below cloud top to 400 m above cloud top (with three upward inversion crossings and two downward), a 10-min straight and level leg in the FT approximately 1000 m above cloud top, and then a profile from this level down to the near surface (140-m altitude). This sampling strategy was designed to provide somewhat representative horizontal and vertical in situ sampling of the PBL and the lower FT, as well as good opportunities for remote sensing of clouds and precipitation.

b. UCL definition

This study focuses upon the nature and occurrence of UCLs over the subtropical northeastern Pacific Ocean. We use G-V in situ cloud and aerosol properties to identify UCLs and to characterize their physical properties. Over the southeastern Pacific (Terai et al. 2014), UCLs were found in all cases of mesoscale open-cellular convection and, in these cases, typically assumed the form of horizontally extensive layers in the upper PBL below the trade inversion. These layers typically consist of patches of clear air and patches of cloudy air. In cloud-free conditions, UCLs were marked by very low concentrations of accumulation-mode aerosol \( N_a \) [defined in Terai et al. (2014)] and here as particles with diameters larger than 0.1 \( \mu \text{m} \), whereas under cloudy conditions, UCLs can be identified by very low cloud droplet concentrations \( N_d \) (droplets with radii larger than 1 \( \mu \text{m} \)). Defining UCLs requires the choice of thresholds, and this is somewhat arbitrary. We note that mean concentrations \( (N_a \text{ or } N_d) \) in the upper PBL during cases of open cells over the southeastern Pacific were generally below 10 \( \text{cm}^{-3} \) (Terai et al. 2014), which is 5–25 times lower than typical mean concentrations found in previous field studies of marine stratocumulus (e.g., Martin et al. 1994; Miles et al. 2000). There have been very few studies that have noted such low particle concentrations; yet, as we shall see, these occur relatively frequently in the SCT. We therefore set a threshold for \( N_a \) or \( N_d \) of 10 \( \text{cm}^{-3} \) to distinguish UCLs from “non UCL” samples.

The procedure for classifying samples is as follows: We begin with G-V aircraft data sampled at 1 Hz, or roughly 150 m along the flight direction. This is not necessarily the native frequency of the measurements, but placing all the measurements for each flight into a single universal time frame allows the use of multiple instruments and criteria to classify air samples. We first classify each 1-Hz sample as (i) clear samples, (ii) cloudy samples, or (iii) samples containing precipitation under subsaturated conditions.

(i) Clear samples must satisfy the following: liquid water content measured using the CDP, \( q_{l,\text{CDP}} \), must be lower than 0.01 \( \text{g kg}^{-1} \), liquid water content measured using the 2DC, \( q_{l,\text{2DC}} \), must be lower than 0.001 \( \text{g kg}^{-1} \), and the total drizzle drop concentration measured with the 2DC, \( N_{2DC} \), must not exceed 1 \( \text{L}^{-1} \).

(ii) Cloudy samples must have \( q_{l,\text{CDP}} > 0.01 \text{ g kg}^{-1} \) and must have relative humidity exceeding 95%.

(iii) Samples containing precipitation under unsaturated conditions have either \( q_{l,\text{CDP}} > 0.01 \text{ g kg}^{-1} \), \( q_{l,\text{2DC}} > 0.001 \text{ g kg}^{-1} \), or \( N_{2DC} > 1 \text{ L}^{-1} \) and must have relative humidity below 95%.

Following classification of samples into cloud, clear, and unsaturated precipitation, each of the cloudy and clear 1-Hz samples is then classified as being either a UCL sample or a non-UCL sample using the following criteria:

(i) To be classified as a clear UCL, the 1-Hz sample must be classified as clear and further must have \( N_a < 10 \text{ cm}^{-3} \).

(ii) To be classified as a cloudy UCL, the 1-Hz sample must be classified as cloud containing and must have \( N_d < 10 \text{ cm}^{-3} \).

Once each 1-Hz sample has been classified, statistical analysis is performed as described in the next section.
3. UCL occurrence statistics

Longitude–height cross sections of UHSAS $N_d$ from RF02 to RF15 (Fig. 1) show that very low concentrations ($< 10 \text{ cm}^{-3}$) commonly occur, especially west of $\sim 135^\circ \text{W}$ where the mean low-cloud cover begins to decrease as the SCT gets underway. It is rare that these layers of very low $N_d$ extend to the surface but instead tend to be concentrated at altitudes of 1–2.5 km, with the deeper cases mostly occurring in the western half of the region. Most flights indicate the presence of UCLs ($N_d < 10 \text{ cm}^{-3}$) extending some distance between California and Hawaii, but there is considerable variability between cases. For example, RF07 shows that UCLs were sampled on every leg over the $\sim 2000$-km distance from 132° to 152°W and were growing in height westward with the growth in the PBL. Interestingly, RF02 and RF03, which show few UCLs, were severely impacted by biomass-burning aerosol from forest fires in British Columbia that entrained into the PBL.

The distribution of cloud droplet concentrations $N_d$ sampled in CSET by the G-V shows that just under 25% of low-cloud samples in the entire campaign have values below $10 \text{ cm}^{-3}$ (Fig. 2). Further, the $N_d$ distribution shows a strong negative skewness, indicating a departure from the approximate lognormality characterizing the upper 75% of the $N_d$ values sampled. This apparent change in the nature of the $N_d$ distribution occurs at $N_d \sim 10 \text{ cm}^{-3}$ and possibly suggests that different processes control the microphysical behavior at low versus high cloud droplet concentrations. This change in distribution behavior also provides additional motivation for defining a UCL threshold for $N_d$ of $10 \text{ cm}^{-3}$.

Using the classification described in the previous section, we analyze clear and cloudy UCLs separately. As Fig. 1 shows, the low-level sampling in CSET extended from 128°W, some distance upstream of the stratocumulus maximum (climatologically at $\sim 135^\circ \text{W}$), to 152°W, close to Hawaii and well into the SCT breakup region. Mean low-cloud cover over this longitudinal range falls from 0.7 to 0.4 (see section 6). Our aim in this study is to examine how UCL occurrence frequency varies both with height and with longitude. To do this, we aggregate 1-Hz flight samples flight by flight into 5° longitude and 150-m altitude bins. The longitude bins are chosen to be sufficiently wide that they encompass an entire flight module (see section 2), providing good vertical statistics in the PBL and lower FT within a single longitude bin. Separately for clear and cloudy samples, we determine the fraction of samples in each flight–longitude–height bin that are classified as UCL samples.

An aircraft flying along a path without backtracking can only provide in situ measurements at a single height for any given longitude. In general, this precludes the measurement of vertically integrated quantities such as cloud cover or UCL cover using in situ aircraft data. For example, the aircraft may happen to be sampling a clear region above clouds when there are clouds below. However, the rigid flight sampling strategy used in CSET, most importantly the use of predetermined flight modules along a predetermined flight track, with leg height choices objectively chosen to ensure good sampling of the PBL and lower FT, means that each flight module can provide useful information on the fraction of samples that are within UCLs. We argue that whereas the problem of estimating cloud cover with in situ G-V data is not well posed, an assessment of the fraction of clear sky and clouds that are UCLs is less biased and is statistically meaningful.

For a first look at vertical UCL statistics, we average the UCL fraction profiles over the CSET flights and longitude bins with sufficient samples to give vertical profiles of the fraction of clear and cloudy samples, respectively, that are diagnosed to be UCLs. To be considered sufficient, there need to be at least 30 samples in a given flight–longitude–height bin to be considered in the average. The frequency of both cloudy and clear UCLs increases with height from 0.5- to 1.5–2-km altitude (Fig. 3). The highest fraction of clear samples classified as UCLs is approximately 20%, occurring at 1.5–1.8-km altitude. Cloudy samples are almost twice as likely to be classified as UCLs compared with clear samples, with values of 30%–50% at 2 km, but the vertical trends for each are quite similar. Above 2 km, the clear UCL fraction falls steadily to $\sim 10\%$ at 3-km altitude. The sampling at altitudes above 3 km was too sparse to produce meaningful statistics. Below 0.5 km, there were few clouds sampled in CSET, and only 1%–2% of clear samples were found to be UCLs. Above 2.0 km, there were an insufficient number of flights sampling clouds to produce statistically representative cloudy UCL fractions.

We next estimate the geographical variation of UCLs between California and Hawaii using the following approach. We seek the conditional probability that the column contains a UCL at some level, assuming maximum vertical overlap. For each flight and 5° longitude bin with a sufficient number of samples, we identify the height bin containing the largest UCL fraction (this is done separately for cloudy and clear UCLs). Then we average these maximum fractions over all the flights to give an estimate of the probability that a given longitude bin contains a UCL somewhere in the column. We accept that the aircraft in situ sampling is sparse and does not sample all levels equally within each longitude bin. Thus, our estimate of mean column UCL fraction is
likely to be biased, but it is not obvious whether this will lead to a high or low bias. Assuming random overlap or maximum-random overlap rather than maximum overlap (e.g., Morcrette and Fouquart 1986) will lead to larger estimates of the column fraction, but we note that for many of the CSET cases, there is a single layer (or two adjacent layers) with UCL fractions substantially higher than any of the other layers. In these cases, there is little difference between estimates assuming random and maximum overlap. On the other hand, the rather large (5° longitude) horizontal distance over which we aggregate aircraft data coupled with the limited number

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Fig. 1. Longitude–height cross sections from CSET flights RF02 to RF15 showing the concentration $N_a$ of aerosol particles larger than 0.1 $\mu$m (UHSAS) from all clear samples. Boxes highlight the presence of UCLs (blue colors). RFs 2, 4, 6, 8, 10, 12, and 14 are outbound (westward) flights, and RFs 3, 5, 7, 9, 11, 13, and 15 are the corresponding return (eastward) flights flown 2 days later. Thus, RF02 and RF03 form a “mission pair” such that the RF03 flight plan was designed to resample approximately 48 h later the air masses observed on RF02 advected using boundary layer trajectories as described in Albrecht (2017, manuscript submitted to Bull. Amer. Meteor. Soc.). The thin vertical lines show the longitudes of those air masses that were successfully sampled on both the outbound and return flights. The mean low-cloud cover decreases from $\sim$70% to the east of 130°W to $\sim$40% to the west of 150°W (see Fig. 13), indicating that the entire transect west of approximately 130°W shown here is essentially in the SCT region.
of samples in some height bins may generate UCL column fraction estimates that are too high. For example, for any given flight–longitude–height bin, we only require thirty 1-Hz samples to constitute sufficient sampling. With an aircraft climb or descent rate of 500 m min$^{-1}$, it takes approximately 20 s to profile through 150-m altitude (i.e., the thickness of our height bins). If all 20 s of this sampling consisted of, for example, clear UCL samples, then the clear UCL fraction for this altitude–longitude bin would be approaching unity based on a single ascent–descent profile, and this would be carried through as the column UCL fraction for the entire longitude bin for that flight. If we increase the number of samples required before a particular bin is considered, then we exclude an increasing number of height bins from the averaging and reduce our reliability of capturing the true vertical profile. To provide a quantitative estimate of the uncertainty, we adjust the number of data points required to include a particular longitude–height bin in the UCL coverage estimate from 20 to 100 and examine how this changes the coverage statistics.

Figure 4 shows the CSET estimated conditional probability (given cloudy or clear columns) of finding a UCL in a given column, plotted as a function of longitude along the CSET sampling region from California to Hawaii. Here, we refer to this probability as the fractional UCL coverage. The coverage uncertainties shown reflect the range of values that we obtain when we vary the number of data points considered sufficient for a longitude–height bin to be included in the coverage estimate. The largest uncertainties are on the coverage of cloudy UCLs as cloud layers are typically very thin and less well sampled than clear air, which constitutes a greater fraction of the column. UCL coverage is only few percent to the east of 130°W, consistent with most prior observations in the region within about 500 km of the California coast where very low concentrations of aerosol and/or cloud droplets are relatively rare. A caveat is that only three of the CSET flights in this region had sufficient sampling to contribute to this average, and so our statistics are rather poor in this region. And whereas there have been some notable studies of very low aerosol and/or cloud droplet concentration clouds in the region east of 130°W (e.g., Hindman et al. 1994; Petters et al. 2006), a survey of the published literature of studies in the near-coastal Californian stratocumulus deck (e.g., Hudson and Xie 1999; Miles et al. 2000; Lu et al. 2007) show that very low CCN conditions are relatively rare. This is consistent with a strong offshore gradient in cloud droplet concentration as seen from satellites, where near-coastal mean values exceed 100 cm$^{-3}$ (Wood et al. 2012). West of 130°W, both clear and cloudy UCL frequencies rise rapidly (Fig. 4) to values in the range of 40%–50% between 130° and 140°W and reach as high as 60% by 150°W. The westward increase for clear UCLs is somewhat less noisy than for cloudy UCLs, which also occur somewhat less frequently than clear UCLs west of 140°W. This is because the overall frequency of clouds decreases westward as the extensive stratocumulus clouds break up (thus making statistics less reliable) but may also reflect a general tendency for clear UCLs to be more horizontally extensive than cloudy UCLs. Thus even as the probability of finding a UCL cloud in any given cloudy sample is higher than it is for the clear
counterparts (Fig. 3), clear UCLs tend to occur somewhere in the column more frequently than cloudy UCLs.

4. In situ properties of UCLs
   
a. Aerosol and cloud properties

   In this section, we use the G-V in situ data to examine vertically resolved characteristics of cloudy and clear UCLs. In clear UCLs, concentrations of aerosol particles with diameters exceeding 0.1 \( \mu m \) \( N_a \) decrease with height from values only slightly smaller than the threshold used to define UCLs (10 cm\(^{-3}\)) in the surface mixed layer below 600 m to values of 2–5 cm\(^{-3}\) in the height range (1–2 km) where most UCLs occur (Fig. 5a). Remarkably, mean cloud droplet concentrations \( N_d \) show very similar vertical behavior up to 2-km altitude. The strong connection between the vertical profiles of \( N_a \) and \( N_d \) is consistent with the hypothesis that clear UCLs essentially result from the evaporation of UCL clouds, which, being low in cloud droplet concentration, results in a very low return of cloud-forming aerosol particles to the clear air. An alternative hypothesis could be that some UCL clouds initially form upon saturation of clear UCLs and the associated activation of a very low concentration of accumulation-mode aerosol. In either case, the question of what drives the very low concentrations in the first place must be addressed.

   Although accumulation-mode aerosol concentrations are, by definition, very low in UCLs, the median concentration of condensation nuclei \( N_{CN} \) (particles with diameters larger than 10 nm) in clear UCLs is 50–90 cm\(^{-3}\) (Fig. 5b), that is, over an order of magnitude higher than \( N_a \). Further, median \( N_{CN} \) in clear UCLs is relatively independent of UCL altitude from the surface up to 2 km. In contrast, median \( N_{CN} \) for all clear samples exceeds 130 cm\(^{-3}\) at all levels, increasing to approximately 200 cm\(^{-3}\) in the lower free troposphere. This indicates that clear UCLs have substantially lower total aerosol concentrations compared with non-UCL clear regions at the same altitude. Previous observational and modeling work has suggested that new particle formation can occur in the upper PBL in regions with open cells and strong precipitation scavenging (Petters et al. 2006; Kazil et al. 2011), yet the evidence presented here suggests that whatever new particle formation is occurring, it is unable to raise total aerosol concentrations to levels that would constitute a substantial PBL aerosol source.

   Clear UCLs have anomalously high mean relative humidity exceeding 80% at all levels and closer to 90% in the most common UCL height range of 1–2 km (Fig. 5b). Clear UCLs at altitudes above 2 km show higher mean condensation nuclei (CN) concentrations and lower mean RH than those layers below this, and some of these UCLs are actually in the FT. Integration of the probability distribution functions for clear samples below 3-km altitude (Fig. 6) show that 60% of UCLs have RH exceeding 85%. In contrast, only 28% of all clear samples have such high RH. This figure also demonstrates that roughly a quarter of clear UCLs sampled in CSET had RH lower than 50%. Since PBL RH rarely falls below 60%, such layers are UCLs in the lower FT. Therefore, UCLs are not solely a PBL phenomenon, although their frequency of occurrence in the lower FT appears to be lower than at the top of the PBL (Fig. 3).

   The frequency of occurrence of UCLs occurring higher up in the FT is estimated from the CSET transit legs (6.5–7-km altitude) that were conducted before and after the low-level sampling. Accounting for the lower density at the transit level to estimate the concentration that these layers would have when compressed adiabatically to 1.5-km altitude, we estimate UCL frequency when subsided to be approximately 15%, which is quite close to the observed frequency of clear UCLs in the lower FT (2–3-km altitude) shown in Fig. 3. Thus, it seems reasonable to conclude that the subsiding FT has the potential to be a contributor to occurrences of UCLs found in the PBL. However, the fraction of cloudy samples classified as UCLs is much higher (30%–40%)
in the upper PBL (Fig. 3). Given this and the tendency for clear UCLs below 3-km altitude to be associated with high RH, it seems likely that many clear UCLs in the PBL originate from the evaporation of UCL clouds in the PBL. In Part II of this paper (O et al. 2018, manuscript submitted to J. Atmos. Sci., hereafter Part II), we use a microphysical model to explore a mechanism for the production of UCL clouds and conclude that collision–coalescence in cumulus updrafts is likely sufficient to explain the observed UCLs in the PBL.

Mean liquid water content in UCL clouds tends to be relatively low compared with typical marine stratocumulus clouds, and the majority of the condensate tends to consist of droplets with radii exceeding 37.5 μm that are measured by the 2DC (Fig. 5c). The total condensate (approximately represented by the sum of the CDP and 2DC qL) does not have strong height dependence and is typically 0.1–0.15 g m⁻³. We use the CDP and 2DC size distributions to estimate the mean effective radii rₑ of UCL clouds and find mean values of 20–25 μm (Fig. 5d). As might be expected given that droplets in the 2DC size range constitute the majority of the condensate, the mean effective radii derived from only the CDP (droplets with radii smaller than 25 μm) are considerably smaller (e.g., Wood 2000), indicating that condensate in small drizzle drops contributes significantly to rₑ in UCL clouds and therefore influences significantly the optical properties of UCL clouds.
b. Turbulence

Clouds in UCLs tend to be much less turbulent than non-UCL clouds in the same altitude range. Figure 7 shows estimates of inertial range turbulence derived every second using windowed (1 Hz) vertical velocity variance $\sigma^2_w$ from the high-frequency (25 Hz) vertical wind measurements. Cloudy UCLs typically have mean $\sigma^2_w$ from 0.015 to 0.03 m$^2$s$^{-2}$, whereas cloudy samples that are non-UCLs have values 3–5 times higher than this. This difference is consistent at all longitudes (not shown). Differences for clear samples are weaker, as are the turbulence levels themselves. Above 1-km altitude, clear UCLs have turbulence levels close to the likely minimum value that can be accurately determined with the aircraft turbulence probe (roughly 0.005 m$^2$s$^{-2}$). In the altitude range 0.5–1.5 km, non-UCL clear regions have turbulence levels roughly 30%–100% higher than UCL clear regions, but above 1.5 km, the differences are smaller. Thus, it may be concluded that both clear and cloudy UCLs are typically significantly more quiescent than non-UCL equivalents at similar levels. There are likely some important consequences of low turbulence levels in UCLs. First, a UCL that mixes with air either above or below would typically result in an increase in aerosol concentration in the layer. Second, introduction of drier air via turbulent mixing into a saturated, cloudy UCL would also tend to result in cloud desiccation. The low turbulence levels observed in UCLs, and the implied suppression of vertical mixing, therefore probably helps to explain their high frequency of occurrence. It may also be indicative of the process by which UCLs are formed in the first place. Theoretical and observational evidence suggests that Cu clouds detrain air preferentially at levels where there is static stability in the environmental profile (Raymond and Blyth 1986; Bretherton and Smolarkiewicz 1989; Perry and Hobbs 1996; Zuidema 1998). If UCL clouds are predominantly formed in this way, then one would expect suppressed turbulence as the rising parcels in the cumulus encountered regions of increased buoyancy, causing deceleration of vertical motion. Visual evidence of a rather laminar appearance of many of the layer clouds observed in CSET is presented in section 5.

5. Case studies

The statistical analysis of the properties of UCLs determined using in situ aircraft observations reveals that both cloudy and clear UCLs are common, especially in the upper parts of the marine PBL above the surface mixed layer. In this section, we combine these observations with aircraft and satellite remote sensing information to provide additional context to highlight properties of UCL clouds that are difficult to ascertain using in situ data alone. We choose three case studies that illustrate different aspects of UCLs and their relationship with mesoscale cloud features characteristic of the SCT. The case studies were chosen because each one has a unique combination of in situ sampling, coupled with remote sensing from satellite or from the G-V.

a. 29 July 2015—Research flight 11

On 29 July 2015, CSET RF11 returned from Hilo, Hawaii, to Sacramento and flew for 11 min (1937–1948...
UTC) through the upper levels of an aggregated cumulus cluster situated at 29°N, 145°W approximately 1500 km to the northwest of Hawaii. Figure 8a shows visible satellite imagery of the cluster indicating the track that the G-V flew through the system. Figs. 8b and 8c show photographs indicating extensive quasi-laminar-layer clouds at a level similar to that of the aircraft (~2750 m). Just before exiting the cluster at 1947 UTC, the G-V climbed slightly to an altitude of 3350 m. The G-V passed through both clear and cloudy UCLs, as indicated by $N_a$ and $N_d$ concentrations (green and dark blue dots, respectively, in Fig. 8d) below 10 cm$^{-3}$. Unfortunately, the UHSAS instrument suffered overheating during the period shown, but total aerosol concentrations $N_{CN}$ sampled in clear layers from 1939 to 1947 UTC after the UHSAS overheated remained very low and in the range 10–30 cm$^{-3}$ (red dots in Fig. 8d), indicating likely clear UCLs associated with the high–relative humidity air. In the dry free-tropospheric air prior to the aircraft overflying the cluster (1936–1937 UTC) and after the overpass (1947–1948 UTC), there was a marked increase in aerosol concentrations, indicating that the source of the sampled UCLs in this case is associated with the humid air in the upper reaches of the cumulus cluster. Turbulence measured by the G-V throughout the entire transect is weak ($\sigma_w^2 < 0.02$ m$^2$ s$^{-2}$), consistent with the statistical analysis of UCL cloud and clear layers shown in Fig. 7.

Fig. 8. Case study of UCL cloud and clear layers sampled by the G-V between 1936 and 1948 UTC during flight RF11 on 29 Jul 2015. (a) Visible satellite image from GOES-15 at 1930 UTC showing aggregated Cu cluster approximately 100 km across along with the aircraft flight track (times marked on track). (b) Photograph from the left side of the aircraft at 1938 UTC showing bright Cu clouds and extensive layer (veil) cloud with gray appearance. (c) Photograph taken during passage through optically thin UCL cloud at 1945 UTC showing two-layered cloud structure at the level of the aircraft. (d) Time series of several variables obtained using in situ G-V measurements during passage through system at 2.75-km altitude. Shown are concentrations of aerosol particles larger than 0.1 μm $N_a$ (green; only available until 1939 UTC because of UHSAS overheating), CN concentration (red dots), $N_d$ (blue dots), RH (orange), liquid water content from the CDP (black line) and from the 2DC (red line), and effective radius (cyan). (e) Lidar backscatter curtain below the aircraft from the HSRL. (f) Radar reflectivity factor curtain from the HCR.
Lidar backscatter imagery from the HSRL below the aircraft (Fig. 8e) indicates the presence of numerous horizontally extensive cloud layers with top altitudes ranging from 1.5 to 2.5 km. Perhaps half of the cloud layers with tops around 1.5–2.0 km fully attenuate the lidar, but many do not. Those layers above 2 km rarely fully attenuate the lidar despite some of them (e.g., those from 1942 to 1944 UTC) being as thick as 1 km. We refer to these thin clouds overlying the cumulus cluster as veil clouds. The G-V passed through the tops of these optically thin cloud layers from 1941 to 1943 UTC and measured condensate consisting of both cloud drops and small drizzle-sized drops (measured with the CDP and 2DC, respectively). Total condensate amounts ranged from 0.04 to 0.1 g m$^{-3}$ (Fig. 8d). Measured $N_d$ in these UCL clouds was typically 1–3 cm$^{-3}$ with effective radii ranging from 20 to 50 $\mu$m (cyan dots in Fig. 8d). Despite relatively low condensate loadings, most of the UCL clouds detected by the lidar here produced detectable radar echoes with reflectivities in the range from $-15$ to $-25$ dBZ (Fig. 8f), consistent with significant condensate loading in the form of drizzle drops. Thus, in this case and in many others not shown here, the optically thin veil clouds mostly exist in UCLs.

A remarkable feature of the optically thin veil clouds sampled in this cluster, given that they contain appreciable drizzle-sized drops, is that many have very sharply defined bases (seen in the HSRL and HCR images and also in the photographs in Fig. 8) with little evidence of virga in most cases other than in some of the most geometrically thick but optically thin layer clouds (e.g., 1943–1944 UTC). This suggests that many of the veil clouds may not simply be decaying remnants of precipitating layers but may be being actively maintained. Evidence for mesoscale lofting of PBL air within aggregated trade cumulus clusters is found in large-eddy simulations (Bretherton and Blossey 2017), where such vertical motions appear to be supported by circulations that develop between the suppressed regions surrounding clusters and the active clusters themselves. How such mesoscale ascent and how the UCL cloud microphysics might each impact cloud cover and longevity of veil clouds are questions worthy of further inquiry.

Although in this case we do not have a G-V profile to provide vertical context, a decoupled PBL structure is evident throughout the sampling period. This is clearest before 1937 UTC and after 1944 UTC when there is a relatively strongly scattering surface mixed layer extending up to 1 km overlain with layers of much lower scattering. In places (e.g., at 2.5-km altitude from 1936 to 1939 UTC and just below 1.5-km altitude between 1944 and 1945 UTC), there are curious horizontally extensive layers of increased lidar scattering. It is possible that these layers are caused by the evaporation of cloud droplets (the scattering layers are horizontally adjacent to cloud layers).

In addition to the qualitative information provided by the lidar and radar sensors on the G-V, a real strength of these instruments is in their combination with each other to produce quantitative retrievals of cloud properties. The radar–lidar sensor synergy is at its best for the optically thin veil clouds, where lidar backscatter is available throughout the cloud layer. Figure 9 shows retrievals of liquid water content and droplet effective radius using a variant of the O’Connor et al. (2005) method (V. Ghate 2018, personal communication) for the same 29 July cloud case shown in Fig. 8. Retrieved liquid water contents ranging up to 0.5 g m$^{-3}$ are found, especially lower down in the cloud layer below the level of the G-V. Effective radii exceeding 20 $\mu$m are found in much of the cloud layer, yielding estimates of cloud droplet concentration $N_d$ that are a few droplets per cubic centimeter and in good agreement with in situ estimates (Fig. 9d). In this study, we show only this one case, but future work could include a systematic analysis of the lidar–radar retrievals from all the CSET flights to help quantify their frequency, liquid water path, and droplet concentration.

### b. 27 July 2015—Research flight 10

On 27 July 2015, CSET RF10 flew from Sacramento to Hilo and sampled large clusters of aggregated cumulus convection and associated veil clouds west of the Sc–Cu transition (138°W). A time-lapse set of GOES visible imagery illustrates the temporal evolution of optically thin veil clouds (reflectances of $\sim0.2$) and demonstrates their relationship with bright (reflectances exceeding 0.4) Cu clouds (Fig. 10). In this case, the G-V flew a straight leg through a stratiform cloud at the top of the PBL (~2300 m). Low cloud droplet concentrations measured with the G-V were in the UCL range during the latter half of the leg (Fig. 11c). Effective radii exceed 20 $\mu$m over much of this leg (Fig. 11), even reaching as high as 40–60 $\mu$m. Moderate Resolution Imaging Spectroradiometer (MODIS)-retrieved cloud droplet effective radii are also very high over a large fraction of the cloud system (Fig. 11b) with independent estimates from the combined CDP–2DC and HOLODEC in excellent agreement (Fig. 11d). An assessment of the utility of HOLODEC to bridge the size gap between the CDP and the 2DC, especially in clouds with a substantial number of droplets in this size range, can be found in Glienke et al. (2017).

The temporal evolution seen in the GOES imagery is particularly revealing, as it demonstrates several key aspects of the life cycle of veil clouds. Active Cu clouds appear to be necessary for the initial formation of the stratiform veil clouds (Cu without veil clouds develop them after 1–3 h as seen in Figs. 10e–h). After the active
Cu weaken and dissipate (e.g., Figs. 10e–g), the veil clouds persist for a few more hours but then eventually evaporate (Figs. 10h,i). In this case, there was relatively little vertical shear of the horizontal wind within the PBL, and the veil cloud locations remain quite close to the Cu that produce them. However, in other cases with directional wind shear in the PBL, veil clouds in the upper PBL can move in a direction opposite to the direction of motion of the Cu, and this sometimes leaves isolated patches of optically thin veil clouds completely disconnected from the Cu that generated them in the first place. Figure 10g also demonstrates that aggregated Cu clusters are associated with rather large increases in column water vapor (from 25 mm in the suppressed regions to as high as 40 mm in the Cu patches). This concentration of moisture in the regions of active Cu appears to be a typical feature of shallow convective self-aggregation, as indicated by large-domain large-eddy simulations (Bretherton and Blossey 2017).

c. 19 July 2015—Research flight 7

Here, we examine data from CSET research flight 7 on 19 July 2015 and compare in situ and satellite-retrieved cloud properties, in this case from GOES-15. Photographic and satellite observations show intermingled shallow cumulus convection and stratiform veil clouds during the G-V traverse through an extensive field of open-cellular clouds on a return leg from Hawaii to Sacramento (Figs. 12a–c). The aircraft remained within clouds with similar open-cellular morphology from 152° to 135°W, a distance of over 1700 km sampled over approximately 4 h (1810–2215 UTC). During this period, the aircraft made numerous cloud penetrations, and GOES-15 cloud product retrievals (using a combination of 0.63- and 3.9-μm-wavelength channels; Minnis et al. 2008a,b, 2011) were processed for the 17 images (15-min temporal resolution) obtained during the aircraft sampling period. Retrievals from all pixels within a 2° swath around the aircraft track were used to produce histograms of cloud droplet concentration (Painemal et al. 2015) and effective radius for the transect that are compared with those derived from the in situ data (Figs. 12d,e). Although there is a rather large-scale mismatch between the aircraft in situ microphysical estimates and those from the rather coarse (~4 km) GOES pixels, the two estimates agree fairly well, with median droplet concentrations \( N_d \) of 7 cm\(^{-3} \) (GOES) and 5 cm\(^{-3} \) (in situ) and median effective radii of 24 μm.
The histograms do show some differences, with larger tails seen in the in situ data, which may be symptomatic of the different spatial scales of the measurements, but in general the agreement between the aircraft and satellite retrievals is rather good. This is somewhat surprising because these open-cellular regions do contain a significant fraction of clouds that are not particularly well suited to visible–near-infrared (VIS–NIR) satellite retrievals, which assume plane-parallel clouds, and may suffer from ambiguity problems for clouds of low optical thickness. Despite this, these findings provide some encouragement that passive satellite retrievals in such aggregated cumulus clusters may not be as poor as has sometimes been supposed, especially if the retrievals are used statistically to characterize the average microphysical properties of cloud systems rather than at the pixel level.

Fig. 10. Evolution of active cumuli and veil clouds on 27 Jul (CSET RF10 flight) seen in GOES visible imagery over a period of 8 h. GOES visible imagery at 1-km resolution over a domain of approximately 3° longitude by 2° latitude that is advected westward with the G-V-observed mean PBL wind. The G-V flew a straight leg through relatively optically thin veil clouds at 2250–2300-m altitude from (d) 1917 to 1927 UTC, where cloud droplet concentration varied from 30 to 40 cm⁻³ during the first one-third of the leg (1917–1920 UTC) before dropping to values in the UCL range 1–10 cm⁻³ after 1922 UTC (Fig. 11). The time-lapse imagery demonstrates that optically thin stratiform veil clouds form in regions adjacent to clusters of optically thick (active) cumulus clouds. Initially, bright Cu form without surrounding stratiform clouds, but (e),(f) veil clouds tend to develop 1–2 h later. When the driving cumulus weaken and dissipate, as occurs after (e) 2030 UTC, (h),(i) the veil clouds persist for some hours before themselves dissipating. (g) An overlay of the column water vapor from the AMSR-2 passive microwave imager on the Global Change Observation Mission–Water (GCOM-W1) satellite that flies in the A-Train. The GCOM-W1 overpass time of 2225 UTC is almost coincident with the 2230 UTC GOES image and shows that the aggregated Cu clusters tend to be associated with quite large moisture perturbations.
The failure rate for VIS–NIR cloud optical property retrievals from NASA’s MODIS is approximately 10% (Cho et al. 2015). The most prevalent type of failure, explaining 60%–85% of all failed retrievals, is that the cloud effective radius implied by the combination of VIS and NIR channels is larger than the upper bound (30 μm) in the lookup table (Cho et al. 2015). The mean UCL cloud effective radius determined using in situ cloud probe data is ~20 μm (Fig. 5d), but the effective radii frequently approach and sometimes exceed 30 μm (see Part II). Given this, the failure of MODIS retrievals due to the presence of large droplets should not be surprising.

Examination of effective radius images from GOES-15 in the region (e.g., Fig. 12b) indicates that extensive regions of the northeastern Pacific are typically covered with clouds with effective radii of 25–30 μm at any one time, a finding consistent with our observations documenting the frequent occurrence of UCL clouds in the region. Thus, it is likely that even if the quantitative accuracy of the GOES-15 retrievals is poor, the estimates may still serve as a useful means to identify and track cloudy ultraclean layers. Further work detailing more extensive comparisons of aircraft and satellite data will be needed to ascertain the extent to which this is the case.

6. Aircraft remote sensing of optically thin clouds

As can be readily seen in Fig. 8, a rather high fraction of the clouds sampled do not fully attenuate the HSRL beam. In this section, we present a statistical analysis to examine how the statistics of optically thin cloud occurrence varies with longitude and height, and this leads to the conclusion that optically thin clouds in the SCT are strongly coupled to the presence of UCLs. The HSRL-determined cloud mask (V. Ghate 2018, personal communication) identifies clouds detected by the HSRL
and also provides a flag indicating whether the beam has been fully attenuated. We use all zenith-pointing data from the near-surface-level legs (see section 2a) to ensure that HSRL has a view of the entire column and therefore can determine the low-cloud cover. For each lidar profile, we first determine if low clouds are present

Fig. 12. Example of extensive patches of thin clouds associated with small cumulus clusters sampled by the G-V for over 1500 km (1750–2120 UTC) on flight RF07 on 19 Jul 2015. (a) A photograph from the left side of the aircraft at 1825 UTC shows a cluster of bright Cu clouds interspersed with horizontally extensive stratiform veil clouds. (b) Satellite-estimated cloud effective radius \( r_e \) from GOES-15 at 1900 UTC (the white symbol on a circular black background shows the location of the photograph). (c) A visible GOES-15 image at higher resolution of the region denoted by the box in (b) showing trade Cu clusters in the form of loosely organized mesoscale open cells. A statistical comparison of histograms of (d) \( N_d \) and (e) \( r_e \) from GOES-15 and from the G-V in situ data taken from the 1500-km portion of the G-V flight indicated by the thick solid line in (b) where the G-V was conducting low-level sampling. See text for details of the data used to construct the histograms.
using the cloud mask. Here, low clouds are defined as those with HSRL-detected bases below 3-km altitude. These data can be used to determine the overall low-cloud cover. Then profiles that contain low cloud are separated into three categories: (i) optically thick low clouds, which are those profiles containing a single low-cloud layer that fully attenuates the lidar beam; (ii) a single layer of optically thin low cloud (i.e., a low cloud that does not fully attenuate the lidar) determined by detecting a lidar signal from above the cloud layer; and (iii) multiple layers of low clouds. This latter category could contain multiple optically thin low-cloud layers or optically thin low-cloud layers below optically thick low-cloud layer above. For the single layer optically thin cloud, we are able to determine both the cloud-layer base and the top and thus the cloud thickness.

Cloud statistics from all CSET flights are aggregated into 5° longitude bins from 125° to 155°W (Fig. 13). Consistent with existing satellite climatologies (e.g., Karlsson et al. 2010; Leahy et al. 2012), low-cloud cover decreases moving westward as the MBL transitions from relatively shallow at the location of the stratocumulus maximum around 130°–135°W to a deeper MBL close to Hawaii. Low-cloud cover decreases as the MBL deepens and the stratocumulus–cumulus transition occurs. The fraction of low clouds that are optically thin increases westward from below 0.4 east of 135°W to 0.6–0.7 west of 140°W. These high values are consistent with CALIPSO lidar-based climatologies (Leahy et al. 2012; Guzman et al. 2017). We further subdivide the optically thin low clouds into those occurring in the upper MBL (bases at or above 1-km altitude) that we refer to as veil clouds $f_{\text{low, thin}, z > 1 \text{km}}$ and those with bases near the lifting condensation level (bases below 700-m altitude $f_{\text{low, thin}, z < 700 \text{m}}$). Figure 13 shows that most of the optically thin low clouds observed in CSET occurred in the upper PBL above 1 km and are therefore considered veil clouds. Optically thin LCL clouds are far less numerous and nowhere contribute more than 15% of all optically thin low clouds. The increasing frequency of optically thin low clouds moving westward mirrors the increase in frequency of both clear and cloudy UCLs (Fig. 4), consistent with other lines of evidence indicating a tight connection between strongly depleted layers and the occurrence of optically thin veil clouds.

The heights and thicknesses of all the optically thin cloud layers detected by the zenith-pointing HSRL during CSET are summarized in Fig. 14. The figure shows box-and-whisker plots showing geometrical thickness versus cloud-base height, with the blue curve providing the cumulative frequency of cloud-base height. The median cloud-base height of the optically thin clouds is 1150 m, with 10th and 90th percentiles of 625 and 1700 m, respectively. Mean geometrical thicknesses of the optically thin layers are just over 200 m (medians are a little lower: ~180 m), with little evidence of a systematic variation of thickness as base height increases. Some optically thick layers are as thick as 300–400 m; the optically thin clouds shown in Fig. 8 exceed this thickness in places. Most optically thin clouds are therefore occurring in the upper PBL above the surface mixed layer, whose top is typically 600–700 m. The height distribution of the fraction of cloudy samples that are UCLs determined using in situ data (Fig. 3) also shows few UCLs in the surface mixed layer and a rapid increase with height, which is consistent with the idea that many optically thin low clouds are also UCL clouds. The red dashed line in Fig. 14 shows the cumulative distribution of the heights of UCL clouds determined using the in situ G-V cloud probe data. It should be stressed, however, that the cumulative frequency distribution of the heights of UCL clouds may be biased by nonuniform height sampling by the G-V. To attempt to mitigate this potential sampling bias, we normalized the number of cloudy UCL samples identified at each level by the total number of aircraft samples at each level. The cumulative distributions of the lidar-determined optically thin cloud-base height and the in situ–determined UCL cloud height are similar, providing additional evidence of a close connection between optically thin clouds and UCL clouds. The two cumulative distributions do diverge somewhat at heights above about 1200 m, with almost twice as many cloudy UCLs occurring above 1500 m than optically thin clouds.

![Figure 13](https://example.com/fig13.png)
7. Discussion

The two key findings in this paper are that (i) ultra-clean layers are a frequent occurrence in the MBL in the stratocumulus-to-cumulus transition region of the northeastern subtropical Pacific Ocean and (ii) many clouds in UCLs tend to be optically thin, quasi-laminar stratiform veil clouds containing large droplets but relatively low liquid water contents. These clouds tend to occur in the upper part of the decoupled trade wind PBL and are only infrequently found in the surface mixed layer below 600 m or so. Veil clouds are particularly common in cases of open mesoscale cellular convection and in large (∼100-km wide) aggregated cumulus clusters. The tendency for a high frequency of optically thin clouds in open- and disorganized cellular convection has been previously observed [see, for example, the cloud optical thickness distributions shown in Fig. 14 in Muhlbauer et al. (2014)], but their microphysical and detailed structural properties have not been previously described. Veil clouds tend to be geometrically as well as optically thin, so their low optical thickness can be attributed to a combination of their thinness and their very low droplet concentrations (optical thickness scales with $N_d^{1/3}$). Veil clouds are relatively free of turbulence and so assume a laminar morphology when viewed from the side. In the heart of the SCT upwind of Hawaii, the aircraft observations suggest that ∼50% of cloudy columns contain UCL clouds, and a similar fraction of low clouds is optically thin in this region, again consistent with the idea that most veil clouds are severely depleted of droplets. This poses the question of how UCL/veil clouds are formed, how they are maintained, how they eventually dissipate, and what aerosol they leave behind upon evaporation. Visible geostationary imagery over a period of several hours demonstrates that veil clouds tend to form 1–2 h after the formation of bright, active Cu cloud clusters and then persist for roughly the same amount of time once the active Cu weaken and disappear. Although we have not conducted an exhaustive study of veil cloud lifetimes, the veil clouds examined here last for several hours in total. Thus, veil clouds are a common feature of aggregated cumulus patches and likely constitute the majority of the cloud cover within these patches.

a. What determines the life cycle and microphysics of UCLs and veil clouds?

In Part II, we use microphysical parcel modeling to explore the idea that UCL clouds are created by rapid removal of cloud droplets by collision–coalescence in the moist updrafts of the active Cu clouds in aggregated clusters or in open cells so that when these Cu detract saturated air, it is already strongly depleted in droplets. This detrainment mechanism for UCL cloud formation appears to capture some key microphysical properties of UCL clouds and can explain their longitudinal variation within the stratocumulus-to-cumulus transition, but other hypotheses may also explain them. For example, if aggregated Cu clusters are regions of weak mesoscale ascent, as is suggested in large-eddy simulations (e.g., Bretherton and Blossey 2017), then it might be possible for UCLs to form in situ rather than from detrainment of condensate from the more active Cu towers. However, if this were the case, then it is difficult to explain how the accumulation-mode aerosol concentrations $N_a$ in clear UCLs can have formed. Slow ascent on the order of ∼1 cm s$^{-1}$ [as found in the simulations of Bretherton and Blossey (2017) in the aggregated Cu clusters] would indeed tend to result in rather low $N_a$ even when $N_a$ exceeds the UCL threshold of 10 cm$^{-3}$. However, such a cloud would also contain a significant number of interstitial unactivated accumulation-mode aerosol that...
would remain in place after the cloud dissipated. Thus, it is difficult to explain the frequent occurrence of clear UCLs with such a formation mechanism. Mesoscale ascent could, by providing condensate, enhance veil cloud longevity compared with a case with no mesoscale ascent and may help to explain their observed lifetimes of several hours. The terminal velocity of cloud droplets increases from 1.2 cm s\(^{-1}\) for droplets with radii of 10 \(\mu\)m to 5 cm s\(^{-1}\) for droplets with radii of 20 \(\mu\)m, indicating that sustained mesoscale ascent of several centimeters per second would be required to maintain a UCL cloud against desiccation by sedimentation of droplets out of the layer. Whether such mesoscale lofting occurs for a sufficiently long time within aggregated cumulus clusters warrants further study using large-domain cloud-resolving models. It is possible that size sorting of droplets occurs in UCL clouds over the cloud life cycle such that the largest droplets are removed by sedimentation out of the cloud base, leaving behind smaller droplets. This temporal evolution of UCL cloud microphysics is explored in Part II.

b. Condensate loading and optical thickness of veil clouds

Based on the measurements presented here, we can make a first attempt to quantify the mean condensate loading contained in veil clouds in the SCT. First, we note the mean veil cloud liquid water contents (sum of the CDP and 2DC liquid water contents) are \(\sim 0.12 \text{ g m}^{-3}\) (Fig. 5c). It is worth noting that an uncertain but likely significant fraction of the condensate is contained in small drizzle-sized drops larger than the CDP upper limit (25-\(\mu\)m radius) but smaller than the first reliable bin of the 2DC (37.5-\(\mu\)m radius), so this condensate loading may be a lower limit. Next, we found that the mean geometrical thickness of veil clouds is just over 200 m (Fig. 14). The product of the mean veil cloud liquid water content and the geometrical thickness yields an estimate of the mean liquid water path (LWP) of approximately 25 g m\(^{-2}\), which is significant, although it should be pointed out that this LWP is close to the detection limit for passive microwave retrievals (Wentz 1997), making accurate spaceborne quantification of condensate loading in veil clouds challenging.

Cloud optical thickness (e.g., George and Wood 2010) is related to LWP and \(N_d\) via

\[
\tau \approx KN_d^{1/3} \text{LWP}^{5/6},
\]

where \(K\) is a weak function of temperature and pressure and \(K \sim 0.25 \text{ kg}^{3/6} \text{m}^3\) for \(T = 280 \text{ K}\) and \(p = 850 \text{ hPa}\). For the typical veil cloud LWP of 25 g m\(^{-2}\) and \(N_d \sim 5 \text{ cm}^{-3}\) (Fig. 5a), \(\tau \sim 2.2\). This low value of \(\tau\) is consistent with the frequent penetration of veil clouds by the lidar. If veil clouds did not have such low \(N_d\) but instead had values close to the accumulation-mode aerosol in the surface mixed layer (\(\sim 75 \text{ cm}^{-3}\); see Part II), then from Eq. (1), \(\tau \sim 5\), in which case, lidar would not penetrate the cloud. Indeed, for a threshold optical depth of 3, below which lidar will penetrate (Winker and Poole 1995), \(N_d\) must be below approximately 20 cm\(^{-3}\). Thus, the low optical thickness of veil clouds is strongly contingent on their low droplet concentrations. Given the frequent occurrence of veil clouds, this implies that there may be a direct control of regional cloud albedo by precipitation through coalescence scavenging.

c. On the radiative susceptibility of veil clouds

CSET data show that many UCL clouds, despite being optically and geometrically thin, have radar echoes significantly higher than \(\sim 30 \text{ dBZ}\) such that they can be detected with the W-band radar on the G-V aircraft and may even be detected and characterized using spaceborne radar. Radar–lidar retrievals designed originally for quantification of the microphysical properties of drizzle show considerable promise for remote characterization of the microphysical properties, condensate loadings, and potentially the albedo susceptibility of veil clouds to changes in aerosol loading. The relatively low albedo \(A\) of veil clouds does not drastically reduce their albedo susceptibility to changes in droplet concentration \(N_d\), which scales as \(dA/dN_d = (1 - A)/3N_d\) (Platnick and Twomey 1994) and thus is only about one-third lower for \(A = 0.2\) compared with \(A = 0.5\).

On the other hand, the extremely low \(N_d\) in UCL clouds implies a very high albedo susceptibility (Twomey effect). Marine low clouds shoulder a large fraction of the global indirect aerosol forcing in climate models (e.g., Kooperman et al. 2012; Carslaw et al. 2013; Lee et al. 2016), so it is important to understand whether models have the appropriate albedo susceptibility given the tendency for large-scale models to produce an insufficient fraction of optically thin clouds. That said, the albedo susceptibility defined by Platnick and Twomey (1994) does not deal with the extent to which \(N_d\) can be perturbed by increases in anthropogenic aerosol emissions. For MBL clouds over the remote oceans, the pathways by which \(N_d\) can be increased are either via increased marine shipping emissions (surface sources) or by the entrainment of long-range transported polluted air from the FT. But in order to perturb veil cloud \(N_d\), surface air must first pass through cumulus clouds to reach the upper MBL. Coalescence scavenging in cumulus updrafts is likely to strongly limit the perturbation to veil cloud \(N_d\) that would be expected from a given increase in aerosol loading in the surface mixed layer.
Thus, although veil clouds have high albedo susceptibility, microphysical processes driving their production may limit their ability to realize this susceptibility. Further, because veil clouds tend to be quiescent, their ability to entrain FT air and tap into the higher aerosol concentrations there may also be strongly limited, too. Exploring the degree to which veil clouds can be perturbed by aerosol is an important topic for further study.

Beyond the Twomey effect, there is the question of whether the colloidal stability of low-\(N_d\) veil clouds may change if aerosol loadings increase. Answering this will require a much greater understanding of the life cycle of veil clouds and of the factors that control their coverage and lifetime. An additional aspect of the susceptibility of the top-of-the-atmosphere radiation to changes in the optical thickness or coverage of veil clouds is that they likely have significant longwave cloud radiative effects by virtue of their temperature contrast with the surface and because emissivity is sensitive to droplet effective radius for typical UCL cloud droplet sizes (Wood 2012).

d. Outlook for future research

Over the global oceans, roughly half of all low clouds do not fully attenuate spaceborne lidar Leahy et al. (2012), indicating the important contribution of optically thin MBL clouds to global cloud cover, especially in the subtropics and tropics. Less is known about the contribution of these clouds to the overall cloud radiative effect. In the SCT regions, these optically thin clouds comprise a mixture of small fair-weather trade cumulus clouds with bases typically tied to the lifting condensation level (LCL) and much more horizontally extensive stratiform veil-type clouds that reside in the upper MBL. Quantifying the contribution of each of these types of optically thin cloud to cloud cover and to overall radiative effect is a significant challenge but one that can probably be largely achieved with current and planned active spaceborne remote sensing technology. Understanding of the large-scale conditions that encourage or discourage UCL production is needed. A major challenge is to accurately represent veil clouds and UCLs in large-scale numerical models and to understand how they may change in response to increasing greenhouse gases and to aerosol changes. Parameterizations that are able to treat cumulus convection and stratiform veil clouds as tightly coupled entities with strongly interacting microphysical and macrophysical processes are likely to bear most fruit.

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