Observations Pertaining to Precipitation within the Northeast Pacific Stratocumulus-to-Cumulus Transition

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(Manuscript received 12 July 2019, in final form 25 December 2019)

ABSTRACT

Three genuine stratocumulus-to-cumulus transitions sampled during the Cloud System Evolution over the Trades (CSET) campaign are documented. The focus is on Lagrangian evolution of in situ precipitation, thought to exceed radar/lidar retrieved values because of Mie scattering. Two of the three initial stratocumulus cases are pristine [cloud droplet number concentrations ($N_d$) of $22 \text{ cm}^{-3}$] but occupied boundary layers of different depths, while the third is polluted ($N_d \sim 225 \text{ cm}^{-3}$). Hourly satellite-derived cloud fraction along Lagrangian trajectories indicate that more quickly deepening boundary layers tend to transition faster, into more intense but more occasional precipitation. These transitions begin either in the morning or late afternoon, suggesting that preceding night processes can precondition or delay the inevitable transition. The precipitation shifts toward larger drop sizes throughout the transition as the boundary layers deepen, with aerosol concentrations only diminishing in two of the three cases. Ultraclean ($N_d < 1 \text{ cm}^{-3}$) cumulus clouds evolved from pristine stratocumulus cloud with unusually high precipitation rates occupying a shallow, well-mixed boundary layer. Results from a simple one-dimensional evaporation model and from radar/lidar retrievals suggest subcloud evaporation likely increases throughout the transition. This, coupled with larger drop sizes capable of lowering the latent cooling profile, facilitates the transition to more surface-driven convection. The coassociation between boundary layer depth and precipitation does not provide definitive conclusions on the isolated effect of precipitation on the pace of the transition. Differences between the initial conditions of the three examples provide opportunities for further modeling studies.

1. Introduction

The transition from overcast stratocumulus to more broken shallow cumulus clouds is a conspicuous feature of all of Earth’s subtropical oceanic basins. The accompanying change in the top-of-the-atmosphere albedo, and the contribution to global hydrologic cycle through evaporation off the ocean’s surface as the boundary layers deepen, has inspired research into the processes underlying the stratocumulus-to-cumulus transition (SCT). The original studies established that the shallow cloud transition is primarily dependent upon the ratio of surface latent heat fluxes to cloud-top longwave radiative cooling (Krueger et al. 1995; Bretherton and Wyant 1997). More recent research has focused on articulating the pace of the transition. One comprehensive analysis of Lagrangian trajectories based on reanalyses and satellite observations concluded that changes in the underlying sea surface temperature, more so than in the atmosphere, dominate the speed of the cloud transition (Sandu et al. 2010).

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DOI: 10.1175/MWR-D-19-0235.1

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Precipitation is also a pronounced feature of shallow subtropical clouds (vanZanten et al. 2005; Burleysen et al. 2013; Mechoso et al. 2014; Zhou et al. 2015; Dzambo et al. 2019), particularly so of the northeast Pacific subtropics compared to the North Atlantic and Southern Hemisphere subtropical regions (Nelson and L’Ecuyer 2018). Precipitation, although a secondary influence on the transition overall (Krueger et al. 1995; Bretherton and Wyant 1997), may help influence the pace of the transition. Paluch and Lenschow (1991) propose that condensational warming from precipitation production, coupled with cooling from precipitation evaporating within the subcloud layer, stabilizes the subcloud and cloud interface, discouraging downward mixing driven by cloud-top longwave radiative cooling. At the same time, evaporation of precipitation creates an instability between the subcloud layer and the surface. This can help clouds remain coupled to the surface, serving to maintain the cloud in stratocumulus regions (Feingold et al. 1996), but in deeper, decoupled boundary layers over warmer waters, the same near-surface instability can aid the surface-driven growth of cumulus clouds. These cumulus clouds can penetrate the stratocumulus deck and enhance cloud-top entrainment that in turn thins the upper-level stratiform layer. In large-eddy-scale simulations, this thermodynamic modification of the stability profile by precipitation can accelerate the cloud transition (Sandu and Stevens 2011). A microphysical hastening of the SCT has also been proposed by Yamaguchi et al. (2017), in which precipitation first develops in cumulus clouds and then is detrained into the stratiform outflow, ultimately removing aerosol from the boundary layer. A similar mechanism is explored within O et al. (2018).

Although modeling studies typically conclude that increased precipitation encourages cloud breakup, observational studies, which have more difficulty distinguishing individual effects, are less conclusive. An evaluation of space-based radar observations of precipitation along Lagrangian trajectories concludes that rain has little influence on the SCT time scale within the northeast Pacific, once the depth of boundary layer and cloud-top inversion strength are accounted for (Eastman and Wood 2016). This could be interpreted to mean that precipitation adapts to the boundary layer depth on time scales of less than a day such that regardless of precipitation, deeper boundary layers transition faster (Stevens et al. 1998; Eastman and Wood 2016). The ship-based Marine ARM GPCI Investigations of Clouds (MAGIC) study also did not find an obvious relationship between precipitation and time scale of cloud breakup in the northeast Pacific, but instead highlighted the importance of a dry free troposphere just above cloud top, which will support cloud-breakup through the entrainment of drier air (Zhou et al. 2015; Mohrmann et al. 2019), in a companion study to ours examining a wide range of cloudiness transitions in the northeast Pacific, did not find consistent changes in precipitation and cloud droplet number concentrations as a function of the SCT.

Modeling studies are optimal for examining effects from precipitation on shallow cloud processes, but model microphysical representations can vary significantly (vanZanten et al. 2005; Wood 2005b; Seifert and Stevens 2010; Li et al. 2015). These differences will directly influence, for example, the distance that a precipitation-sized drop can fall while evaporating in the subcloud layer, affecting the magnitude of the net atmospheric diabatic heating, its vertical structure, and thereby its ability to form atmospheric cold pools and help organize further convection. Our study, motivated by the aforementioned processes, documents the highly detailed precipitation observations collected during the Cloud System Evolution in the Trades (CSET; Albrecht et al. 2019) campaign and examines how precipitation may influence the SCT. The CSET campaign, held during July–August 2015 in the northeast Pacific, benefitted from a long-range research aircraft equipped with new remote sensors, and a novel sampling strategy in which air parcels originally sampled during California-to-Hawaii (CA-to-HI) flights were resampled approximately 2 days later, after the clouds had fully transitioned to cumulus clouds, with HYSPLIT Lagrangian trajectories identifying the location of the evolving air mass. The CSET goal of implementing a true Lagrangian sampling allows microphysical changes to be more readily related to the vertical structure of the environment, as can be done in a model. Aircraft in situ studies of SCT are scarce, reflecting the logistical challenge of Lagrangian sampling over a large domain with platforms moving at a different characteristic speed than that of the clouds.

A guiding question is whether a relationship can be determined between the pace of SCT and the presence and intensity of precipitation. A further driving question is how the precipitation affects latent heating and its vertical profile, and how that changes as a function of SCT. This study complements the more comprehensive assessment of all the tracked Lagrangian evolutions available within Mohrmann et al. (2019). The constraints placed on the flight domain meant that most of the initialization locations did not originate in stratocumulus cloud (e.g., Fig. 6 of
Albrecht et al. (2019) but rather within boundary layer cloud already undergoing transition. The present study selects for those Lagrangian trajectories containing true stratocumulus-to-cumulus transitions, with stratocumulus cloud identified as single-layered overcast clouds using cloud radar reflectivities during the upward-looking subcloud legs. This focus allows us to identify the transition to more broken cumulus using the hourly cloud fractions derived using infrared brightness temperatures from a geostationary satellite. In so doing the cases that are best suited for follow-up modeling studies are selected. Although the small sample size will not support statistically significant findings, the in situ data provide opportunity to characterize the transitions to greater microphysical detail than is possible with reanalyses and remote sensing datasets.

Section 2 describes the CSET campaign and datasets. In section 3 we put forth the change in sea surface temperature (SST) and cloud characteristics along the parcel Lagrangian trajectories. The HYSPLIT model trajectories are assessed in section 4 using carbon monoxide (CO) mixing ratio profiles, and corresponding thermodynamic and moisture profiles are discussed. Section 5 examines in situ precipitation characteristics at the beginning and endpoints of the trajectories. The cloud fraction changes along the transition are quantified using hourly GOES-15 data in section 6, with a “transition time” determined. The low-level precipitation changes are related to changes in the near-surface relative humidity and cloud base in section 7, extended with results from a simple subcloud evaporation model. Radar–lidar rain-rate retrievals are evaluated using the in situ microphysics in section 7 as well, ending with further discussion in section 8.

2. Cloud System Evolution over the Trades: Campaign description and datasets

The CSET campaign began on 7 July 2015 with a flight from California to Hawaii and the last flight back from Hawaii to California occurred on 9 August 2015. The flight numbers, dates, and directions of all the individual flights are provided in Table 1. Each flight included a sequence of three to six boundary layer modules, with ferry legs bookending both ends of each flight. The ferry legs, flown at an elevation of 6–7 km, comprise approximately 30% of each flight, and include dropsondes deployed approximately every 28 in longitude. A typical boundary layer module consists of three 10-min horizontal level legs, one at an altitude of approximately 150 m, another an in-cloud (near cloud base) level leg, and the third an above-cloud level optimized for the cloud remote sensing. The module also includes a 10-min segment of ascending/descending profiles through the cloud (“porpoise legs”). The average flight speed of 130 m s−1 in the subcloud layer corresponds to a distance of almost 80 km over 10 min. A full boundary layer module takes 40–50 min, spanning a distance of 400–500 km. The boundary layer modules are hereafter labeled by the flight number, typically preceded by RF (Research Flight) followed by an alphabetical lettering of the modules (“a,” “b,” etc.), beginning with the easternmost module. The full complement contained within each boundary layer module allows the in situ microphysical measurements to be related to the cloud radar and lidar measurements and to
changes in the inversion and cloud base height, determined relatively nearby.

The climatological maximum stratocumulus cloud fraction in the northeast Pacific is at 20°–30°N, 120°–130°W (Klein and Hartmann 1993) with Lagrangian trajectories emanating from there previously examined in Sandu et al. (2010). The two aircraft deployment locations of Sacramento, California (CA), and Kona, Hawaii (HI), combined with a flight time not to exceed eight hours while maintaining the objective of a Lagrangian resampling 2 days later, constrains the “outbound” CA-to-HI flight path to a more northerly route. This route initially follows along 40°N, with the first boundary layer module beginning at 40°N, 130°W, or 10°N and at the western edge of the climatological maximum. Forward trajectories, calculated using HYSPLIT (Stein et al. 2015) and the National Centers for Environmental Prediction (NCEP) Global Forecast System meteorology, were initialized at 40°N, 130°W and further west at approximately 0.5° intervals and maintained a constant height of 500 m (Mohrmann et al. 2019). The endpoints of these Lagrangian trajectories were then resampled during the HI-to-CA “inbound” flight. The trajectories were redone after the flight using NCEP reanalysis winds to better identify the true Lagrangian endpoint. Further details on the complete Lagrangian trajectories and their assessment are available in Mohrmann et al. (2019).

The 17–19 July (RF06-RF07) flight pair serves as an example of the CSET strategy (Fig. 1). On the “outbound” flight, a stratocumulus deck reaching 600–800 m in height is first sampled during module RF06a at 37°–40°N, 130°–135°W. The initial descent from the ferry leg to near the surface fully profiles the lower free troposphere, followed by a level leg that characterizes the subcloud layer. The subsequent 10-min level leg at approximately 600 m captures the cloud base cloud/precipitation microphysics, followed by a profiling of the cloud microphysical and thermodynamic vertical structure, then a 10-min in-cloud level leg and an above-cloud leg optimized for the remote sensing instruments (Fig. 1b). This is followed by another boundary layer module (RF06b, not shown), with seven Lagrangian forward trajectories initialized within the two boundary layer modules. These diverge into five locations spanning approximately 2000 km total 2 days later, with each endpoint falling within a distinct boundary layer module for this case. The flight track and radar reflectivities are shown for one of the “incoming” modules (RF07b; Fig. 1c). The SSTs (from ERA-Interim reanalysis throughout) have increased toward the equator by approximately 4° (Fig. 1a) and the accompanying clouds have deepened to heights exceeding 1500 m (Fig. 1c).

### a. CSET in situ measurements

Precipitation is primarily characterized using microphysical probe data. This is motivated by a concern that Mie scattering by larger precipitation particles, at the cloud radar wavelength of 3.22 mm (94-GHz frequency), will depress the radar reflectivity measurements, leading to underestimates in rain rates retrieved using conventional radar reflectivity–rain rate relationships. This is not a trivial concern, and is supported by Mie effects noted in Fig. 2 of Schwartz et al. (2019) for a portion of RF07, a flight that figures prominently within the current study. More than 90% of the rain rates measured during CSET are obtained from rain drops greater than 1 mm (Fig. 11). For a 94-GHz radar, the Mie effect becomes apparent when drop diameters...
exceed 200 μm (e.g., O’Connor et al. 2005), CSET radar-based precipitation estimates might be largely affected by Mie scattering. An assessment of the radar-lidar rain-rate retrieval for RF06a and RF07c is also included in section 6.

In situ precipitation is calculated from Two-Dimensional Cloud (2DC) optical array probe measurements. The probe samples drops every second across the 75–3200-μm diameter range at 25-μm resolution (126 diameter bins total, variable C2DCA_LWOO); we note that Wood et al. (2018) used a preliminary version of the 2DC data spanning a more limited diameter range from 87.5 to 1587.5 μm. Rain frequencies and mean rain rates over the 10-min 150-m and in-cloud legs are calculated from 1-s rain rates exceeding 0.01 mm h\(^{-1}\).

A concern remains that in situ sampling may be undersampling the largest drop-sizes (e.g., Wood 2005a). This is addressed by examining the characteristics of the in situ precipitation for all of the flights as a function of SST (Fig. 2). This indicates that rain can be more frequent at the colder sea surface temperatures (<296 K) of the stratocumulus region, than over warmer oceans (Fig. 2). Over waters warmer than 296 K, the rain rates increase, with several 10-min mean values exceeding 5 mm h\(^{-1}\). Overall, the plausible depiction of expected changes of precipitation with sea surface temperature lends confidence that changes in in situ precipitation between the beginning and ending of a Lagrangian airmass trajectory can be meaningfully related to other characteristics of SCT.

Ancillary datasets include cloud droplet number concentrations \(N_d\) calculated from Cloud Droplet Probe (CDP) measurements. The leg-mean \(N_d\) is calculated from 1-s \(N_d\) values > 0. Carbon monoxide (CO) gas concentrations are used to assess the Lagrangian trajectory calculations. CO has an atmospheric lifetime of 1–2 months, with the primary sink being a reaction with OH to form CO\(_2\), and CO mixing ratios will remain conserved over the span of 2 days if no mixing with air containing a different CO mixing ratio occurs. CO mixing ratios of 50 ppbv are typical of the pristine southeast Pacific marine boundary layer (Mechoso et al. 2014), although minimum values of 70 ppbv may be more typical for boundary layers neighboring continents with biomass burning (Diamond et al. 2018). CO mixing ratios are derived from AeroLaser Vacuum Ultra Violet resonance fluorescence measurements at 1-Hz resolution with ±3 ppbv error.

The decoupling and moisture stratification of the boundary layer, strength of the inversion layer, and
free-tropospheric temperature and moisture content are characterized using in situ aircraft profiles of the state variables. Inversion height is set to the maximum of vertical gradient in potential temperature ($\partial \theta / \partial z$) above the lifting condensation level (LCL) (Stull 1988; Oke 1988; Seidel et al. 2010; Zhou et al. 2015), with the additional requirement that $\partial \theta / \partial z$ exceed 0.5 K (100 m)$^{-1}$. The vertical profiles are smoothed using a 15-s moving average. This may place the inversion height slightly above its base. Calculations of the equivalent potential temperature and total water mixing ratio profiles also incorporate the aircraft measurements of cloud liquid water content.

b. CSET remote sensing measurements

The aircraft-borne remote sensors include airborne HIAPER Cloud Radar (HCR) and High Spectral Resolution Lidar (HSRL). HCR is a 94-GHz frequency (W-band) Doppler radar placed in a special pod that allows the radar to point either downward or upward (but not simultaneously). The radar sensitivity is $-39.6$ dBZ at a range of 1 km. The 532-nm wavelength HSRL can also be oriented upward or downward, with a 4° zenith angle offset to minimize specular reflection. The HSRL beam is severely attenuated in cloud and drizzle columns, making it well-suited to sense the near-lidar cloud boundary. This is particularly useful for cloud base, for which precipitation will mask the cloud near-lidar cloud boundary. This is particularly useful for clouds only invalidated this approach occurring between neighboring 1-m levels is thereafter calculated directly from the difference in RWC between the two levels, multiplied by latent heat of vaporization ($L_v$) and terminal velocity:

\[ \text{RWC} = \sum_{i=3}^{i=129} \text{RWC}_i = \sum_{i=3}^{i=129} 4\pi (\rho_i N_i r_i^3) / 3, \]

where $\rho_i$ is the density of water and $N_i$ is that bin’s raindrop-size number concentration. The evaporation occurring between neighboring 1-m levels is thereby calculated directly from the difference in RWC between the two levels, multiplied by latent heat of vaporization ($L_v$) and terminal velocity:

\[ \frac{dr_i}{dh} = \frac{S - 1}{r_i \times D_f \times v_i f_v} \]

(Rogers and Yau 1989), where $h$ is the distance below cloud base, $S$ is relative humidity, $D_f$ is a heat conduction and vapor diffusion constant, and $f_v$ is the ventilation coefficient for vapor transfer. The terminal velocity $v_i$ is calculated using:

\[ v_i = \exp(5.984 + 0.8515 x_i - 0.1554 x_i^2 - 0.03274 x_i^3), \]

where $x_i = \ln[2r_i (\text{mm})]$ (Fang et al. 2017). The relative humidity is assumed to decrease linearly from 100% at cloud base toward the mean value of nonrainy portions measured during the nearest 150-m level leg. The latter was done to emulate the cloud impact on an otherwise undisturbed subcloud layer, but could lead to an overestimate in the evaporation. A new raindrop size distribution is determined every meter. The rainwater content (RWC) is then calculated from the sum over all of the individual bins:

\[ \text{RWC} = \sum_{i=3}^{i=129} \text{RWC}_i = \sum_{i=3}^{i=129} 4\pi (\rho_i N_i r_i^3) / 3, \]

where $\rho_i$ is the density of water and $N_i$ is that bin’s raindrop-size number concentration. The evaporation occurring between neighboring 1-m levels is thereby calculated directly from the difference in RWC between the two levels, multiplied by latent heat of vaporization ($L_v$) and terminal velocity:
The rainwater content at cloud base is set equal to that from the in-cloud leg. Although collision–coalescence will increase the rainwater content between in-cloud leg and cloud base, the assumption that the raindrop size distribution measured during in-cloud leg scales well with cloud-base precipitation is supported by observations (Wood 2005a). The mean cloud base height is derived from lidar measurements from the nearest 150-m level leg. This excludes thin clouds near the inversion base using a threshold based on visual inspection of imagery. We also assume that raindrop-size number concentration remains constant within each bin until those drops evaporate completely. The most limiting assumption may be the neglect of mesoscale variability (e.g., Comstock et al. 2004), including in the subcloud humidity field. However, a large-scale meteorological analysis is consistent with the Lagrangian trajectory flow as well as a reducing inversion strength toward the equator. The main reason for this exercise is simply to develop an intuition of how the subcloud evaporation evolves, and in particular if it increases or decreases over the course of a Lagrangian trajectory, based on mean observed conditions.

e. Selection of Lagrangian trajectories

Eleven boundary layer modules (out of a total of 54 modules) from three flight pairs (out of seven) satisfy our selection criteria for SCT and do not include high clouds. These are connected through 10 Lagrangian trajectories, one of which extends through two boundary layer modules, and are summarized in Table 2. The SCT is most clearly sampled from 17 to 19 July (RF06-RF07), with seven trajectory pathways connecting the two aircraft flight paths, shown in Fig. 1 superimposed on the mean 3-day SSTs, with the boundary layer modules labeled. The other two SCTs occurred on 27 to 29 July (RF10-RF11) and 7–9 August (RF14-RF15).

f. Meteorological context

The average center of the northeast Pacific subtropical high was at about 43°N, 148°W during CSET, displaced slightly north of the climatological position (Albrecht et al. 2019). Sea surface temperatures were ~0.5°C above the 1980–2010 values, which should facilitate the SCT, all else equal (Sandu et al. 2010). The subtropical high is modulated by synoptic activity that will also influence the motion of the Lagrangian trajectories. The synoptic modulation is shown for the three flight pairs using ERA-Interim geopotential heights at 500 hPa (Fig. 3). A strong ridge at around 45°N on 17 July corresponds to a strengthening of the sea level pressure (not shown), reflected in a complete absence of low cloud at that location in the GOES satellite visible imagery (Fig. 5). The strengthening in subsidence is apparent in a lower boundary layer height for the stratocumulus cloud sampled on this day, relative to other flights (shown later). By 19 July the weakening ridge, in combination with a closed low at 25°N, 135°W, can explain the strong divergence evident between the different Lagrangian trajectories. On 27 July, a strong closed trough north of 40°N coupled with a weak ridge to the southeast explains the westward-moving HYSPLIT trajectories of the next 2 days. On 7 August a weaker trough further north of 40°N, combined with a closed high at 20°N, 140°W, also encourages a more westward flow than on 17–19 July.

3. SST and cloud characteristics along the transition

The Lagrangian trajectories emanating from stratocumulus mostly originate at SSTs < 296 K (Fig. 2). The SSTs underlying the air masses for 13 stratocumulus
modules have increased by approximately 5 K by the time the same air masses are resampled during the Hawaii-to-California flights (Fig. 4). GOES-15 visible imagery during the outbound (CA-to-HI) and incoming (HI-to-CA) flights, superimposed with the GV flight tracks, provide large-scale context for the three flight pairs that sample a true SCT and were unobscured by high clouds: 17–19 July (RF06-RF07; Fig. 5), 27–29 July (RF10-RF11; Fig. 6), and 7–9 August (RF14-RF15; Fig. 7).

The three outbound flights sampled boundary layers of different depths with different precipitation characteristics. The lidar-derived cloud base heights and the in situ inferred lifting condensation levels (LCL) from the subcloud layers provide information on how well the clouds are coupled to the surface, with differences of less than 150 m considered well-coupled (Jones et al. 2011). The most consistently precipitating stratocumulus cloud occurred on 17 July (RF06) and spanned two neighboring, similar boundary layer modules (RF6a and RF6b; Fig. 5, middle panel). The cloud radar reflectivities indicate several cells within each leg’s span of ~70 km, with cloud-top heights oscillating around 800 m, and remain well-coupled despite the precipitation. The stratocumulus clouds sampled on 27 July (Fig. 6, RF10a and RF10b) are located higher, with cloud tops at ~1200 m, and with precipitation from a broader cell during RF10a clearly reaching the ocean surface, when the boundary layer becomes decoupled. The stratocumulus cloud on 7 August (Fig. 7, RF14b) also reached cloud-top heights of ~1200 m, and is decoupled but with little precipitation.

Approximately 56 h later, all of the stratocumulus have evolved into cumulus clouds reaching 2 km at times, within overall more clear sky. An interesting...
aspect of the 17 July case is that the Lagrangian trajectories, originally located within 900 km of each other, have diverged as they move to the southwest into five boundary layer modules (RF7a-e) encompassing 2000 km, with the higher cloud-top heights to the west consistent with reduced large-scale subsidence (Mohrmann et al. 2019). In contrast, the trajectories move primarily to the west for the two later flight pairs, and those beginning on 27 July ultimately overlie each other. The cloud structures after 2 days of advection include narrow cumulus towers with low radar reflectivities (Fig. 5, RF7c; Fig. 7, RF15d), intense precipitation inferred from high cloud radar reflectivities (Fig. 5, RF7c), thin solitary stratiform clouds, likely indicating detraining from a parent cloud that is no longer visible or away from the aircraft track (Fig. 5, RF7a-c), and multilayered detraining (Figs. 5 and 6; RF11b and RF15c). The 7–9 August

Fig. 5. (top left) 1700 UTC 17 Jul GOES visible image with RF06a and RF06b modules indicated along with Lagrangian trajectories (green). (top right) 1700 UTC 19 Jul GOES visible image with RF07a, RF07b, RF07c, RF07d, and RF06e modules indicated along with Lagrangian trajectories (green). (middle) Cloud radar reflectivities sampled during the 150-m legs of (left) RF06a and (right) RF06b. (bottom five panels) Cloud radar reflectivities sampled during the 150-m legs of RF07a-RF07e modules. Cloud bases, lifting condensation level (LCL), and flight path indicated by black, green, and cyan lines, respectively. Mean leg ERA-Interim SST indicated in each subplot. Each leg is approximately 78 km long.
flight pair indicates the least precipitation of the three cases.

4. Boundary layer characteristics before and after the cloud transition

The vertical thermodynamic profiles of potential temperature (\(\theta\)), equivalent potential temperature (\(\theta_e\)), and total water mixing ratio (\(q_T\)), constructed from the aircraft ascent and descent datasets, indicate a near-surface \(\theta\) of approximately 290–292 K and water vapor mixing ratios (\(q_v\)) of 10–11 g kg\(^{-1}\) for all of the strato-cumulus regions (Fig. 8). The near-surface \(\theta\) warms to ~296 K and the near-surface \(q_v\) moistens up to 15 g kg\(^{-1}\) in 2 days (Fig. 8a). The inversion heights, marked by full circles, indicate a deepening from a stratocumulus boundary layer depth 960–1280 m with a strong inversion, to inversion depths of 1550–3200 m with less well-defined structures after 2 days. These inversion heights are approximately 100–200 m above the cumulus cloud tops shown in Figs. 4–6. A consistent east–west deepening of the boundary layer is evident on 19 July, from 1600 (RF7a) to 2500 m (RF7c).

The thermodynamic profiles suggest mildly decoupled conditions for all the stratocumulus boundary layer legs, when evaluated using the Jones et al. (2011) threshold criterion of 0.5 g kg\(^{-1}\) in \(q_T\) and 0.5 K in \(\theta\) for the differences between the top and bottom quarters of the cloudy boundary layer. The \(q_T\) differences are 1.1, 0.7, 2.3, 1.6, and 1.5 g kg\(^{-1}\) respectively for RF6a, RF6b, RF10a, RF10b, and RF14b. This suggestion of decoupling contrasts with the conclusions drawn from comparing the time series of cloud base height to the lifting condensation level (Figs. 5–7), also shown in Fig. 5 of Bretherton et al. (2019). The LCL-cloud base comparisons draw on larger sample sizes, suggesting they are more robust. To add context, the thermodynamic profile for RF6b went through clear air, and a
significantly drier, warmer upper boundary layer. The RF10a thermodynamic profile sampled a moister surface mixed layer that is clearly separated from an upper, drier, lower-$\theta_e$ cloud layer. This is consistent with evaporation of some of the precipitation reaching the surface (Fig. 6), also evident in a cooler potential temperature. Such cold pool profiles in stratocumulus regions are also documented in Jensen et al. (2000), vanZanten et al. (2005) and Terai and Wood (2013), even if their mesoscale signatures can be difficult to detect within visible satellite imagery. After the transition, all the thermodynamic profiles show a completely decoupled structure with a decreasing $\theta_e$ and $q_T$ from the surface to the inversion base, indicating no cumulus-coupled conditions were sampled during the aircraft ascents and descents. Although more intense precipitation facilitates the formation of cold pools, freshly formed cold pools do not appear to be sampled after the transition.

The CO mixing ratios (Fig. 8d) are well mixed throughout the boundary layers in both the stratocumulus and cumulus region for most flight pairs, consistent with CO’s long lifetime and imperviousness to cloud processing and precipitation. Once well mixed, only further mixing with air masses of a different CO mixing ratio can alter the vertical structure. Boundary layer CO mixing ratios are within 66–78 ppbv (73–78 ppbv for 6a to 7a-c, 66–72 ppbv for 6b to 7d, 7e, and 69–75 ppbv for 14b to 15c-d) for the 17–19 July and 7–9 August flight pairs. The variations are less than the measurement error. The 27–29 July flight pair is an outlier with boundary layer CO mixing ratios exceeding 110 ppbv, capped by freetropospheric values that range from 60 to about 300 ppbv. The elevated values reflect the influence of biomass burning smoke from fires in North America, with the profiles indicating a general mixing of the smoke from the free troposphere into the boundary layer over the course of the SCT; the free troposphere itself becomes cleaner, presumably through diffusion. The 500-m CO mixing ratios increase by 15 ppbv at the most during the transition. A high overall correlation within
the outgoing and incoming flight pairs of 0.93 based on all of the Lagrangian trajectories, shown in Fig. 3 of Mohrmann et al. (2019) and a correlation of 1 for the 11 SCT Lagrangian pairs in the present study lends a general credibility to the CSET Lagrangian sampling strategy, if difficult to assess independently for the more pristine conditions.

5. Evolution of the in situ precipitation

Changes in the precipitation frequency and rain rates, based on the conditional sampling of 1-Hz rain rates $> 0.01$ mm h$^{-1}$, between the beginning and end points of the Lagrangian trajectories (Fig. 9) approximately follow those shown in Fig. 2 as a function of sea surface temperature. The rain frequencies decrease for all of the trajectories, while the mean rain rates increase for most of the cases (Fig. 9). The distribution of the 1-Hz rain rates within the in-cloud and 150-m legs indicates important differences between the three flight pairs (Fig. 10). The 17 July stratocumulus (RF6a, RF6b) is distinguished by precipitation that likely reached the surface, with a mean stratocumulus near-surface rain rate of $1.2$ mm h$^{-1}$ for the samples deemed raining, or leg-mean rain rates (rain rate when precipitating) of $0.54$ and $0.37$ mm h$^{-1}$ (equivalent to $13$ and $9$ mm day$^{-1}$), for the two boundary layer legs. This is substantially more than the $0.5$–$2$ mm day$^{-1}$ documented by vanZanten et al. (2005) to the southeast of the CSET location during a nocturnal stratocumulus campaign, while Zhou et al. (2015) document little stratocumulus precipitation reaching the surface. Two days

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**Fig. 8.** Vertical profiles of (a)–(d) potential temperature (K), (e)–(h) equivalent potential temperature (K), (i)–(l) total water mixing ratio (g kg$^{-1}$), and (m)–(p) carbon monoxide (CO) mixing ratio (ppbv) for the Lagrangian pairs during (a),(e),(i),(m) 6a to 7a, 7b, 7c; (b),(f),(j),(n) 6b to 7d, 7e; (c),(g),(k),(o) 10a, 10b to 11b, 11c, and (d),(h),(l),(p) 14b to 15c, 15d. The bold filled dots on each vertical profile lines represent the inversion base height.
later, the cloudy air mass is still precipitating, with 20%–25% of the 1-Hz values exceeding 10 mm h⁻¹ on 19 July, indicating intense showers. Leg-mean near-surface rain rates increase less substantially, to 0.55, 2.2, and 1.2 mm h⁻¹ (13, 53, and 29 mm day⁻¹, respectively) within RF7b-d. Near-surface stratocumulus precipitation on 27 July (RF10a) is less substantial but still pronounced for stratocumulus, at 2.6 mm day⁻¹, evolving to 1.5 mm day⁻¹ by 29 July (11b).

One explanation for the more intense stratocumulus precipitation may be the more remotely sampled stratocumulus away from the main stratocumulus deck, with the Second Dynamics and Chemistry of Marine Stratocumulus field study (DYCOMS-II) sampling primarily near 30°N, 123°W. If so, this is further evidence that precipitation can be integral to low cloud evolution. The stratocumulus rain rates documented for these 2 days also exceed those documented for nocturnal stratocumulus in the southeast Pacific (Mechoso et al. 2014; Wyant et al. 2015).

The rain rates from shallow cumuli 2 days later, with depths of approximately 2 km, are somewhat less than those documented during the Rain in Cumulus over Ocean (RICO) campaign in the Caribbean (Rauber et al. 2007; Geoffroy et al. 2014). This is consistent with Caribbean cumuli attaining higher cloud-tops heights, exceeding 3 km (Snodgrass et al. 2009; Zuidema et al. 2012). The CSET precipitation observations are therefore useful for further documenting the relationship of precipitation to boundary layer depth over a wider range of boundary layer depths.

In contrast, no precipitation reaches the surface at either the beginning or endpoint of the more polluted 7–9 August transition. The precipitation differences between the three flight pairs reflect their aerosol/cloud droplet number concentrations. The leg-mean cloud droplet number concentration ($N_d$) of the most polluted stratocumulus cloud, on 27 July, is 224 cm⁻³ (RF10a). This nevertheless precipitating cloud (Figs. 9 and 10) evolves into cleaner cumuli clouds with $N_d$ of 22 cm⁻³ 2 days later. Given that the CO mixing ratio simultaneously increase, this indicates that the aerosols are also entrained, with precipitation responsible for an overcompensating $N_d$ reduction. Truly aerosol-deprived conditions only occurs on 19 July, in which an initially lower stratocumulus $N_d$ of 14 and 25 cm⁻³ on 17 July evolves to leg-mean $N_d$ values of 0.1, 2, 0.1, 2.6, and 1.2 cm⁻³ (based on the cloudy samples only of the in-cloud level legs) for the five boundary layer modules of RF07. An important difference from the more polluted boundary layer of 27 July is that the initial stratocumulus cloud sampled on 17 July was already pristine. This notable cleansing of the boundary layer by precipitation, in which subcloud rain rates on 19 July exceeded 10 mm h⁻¹ at times (Fig. 10), is documented in Wood et al. (2018) and O et al. (2018). A further contrast is provided by the stratocumulus cloud sampled on 7 August during RF14b, in which a similar leg-mean stratocumulus $N_d$ of 22 cm⁻³ did not correspond with rain reaching the surface. The cloud droplet number concentration remained constant, measured to be 21 cm⁻³ during RF15d 2 days later. The difference here is a higher boundary layer depth for the same stratocumulus cloud thickness, lifting the cloud base (Fig. 7) and facilitating an earlier decoupling of the cloud layer from its surface moisture source. Together these three transitions provide a diversity in initial aerosol
and inversion height conditions that lend them well to further exploration through dedicated modeling studies.

The bulk of rain both near the surface and in-cloud is provided by the larger drops (Fig. 11) in both stratocumulus and cumulus regions. More than 90% of the rain rates come from raindrops larger than 1 mm for all the SCT legs. The raindrop size distribution (Fig. 12) indicates a clear shift toward larger drop sizes as the stratocumulus cloud transitions to cumulus. This transition is consistent with collision–coalescence within deeper clouds (e.g., Hudson et al. 2015), with the increased width possibly also reflecting the influence of mixing with environmental air (e.g., Igel and van den

![Figure 10](image_url)
Heever 2017). The concentration of small drops is higher within the cloud than 150-m level. These distributions also allow for an assessment of truncated lognormal and exponential fits. A lognormal fit is used within the rain-rate retrievals developed within Wood (2005b), but the author concludes that both functions provide reasonable fits. We find that the truncated lognormal function provides a superior representation of the rain size distribution in most cases. The exception is when only a few large drops are present. Median values of the diameter, calculated from the lognormal fit, increase from 0.1 mm to 0.2–0.5 mm along the transition, and the geometrical standard deviation also increase. Details of the fitting are discussed in the appendix. These fits may help with assessing common model microphysical representations applied to further study these transitions.

6. Cloud transition from GOES

Time series of the hourly GOES 11-μm infrared-derived cloud fraction, centered on the Lagrangian track, are used to quantify the pace of cloud transitions (Fig. 13). The time of cloud transition is subjectively assigned to the beginning of a 5-h period with consistent cloud fractions below 0.5. A noticeable feature is that the beginning of all the transition cases occur during daylight hours. The broad relationship to the diurnal

![Fig. 11. The 1-s rain-rate cumulative frequency plotted as a function of raindrop diameter for (a) stratocumulus legs-6a-b, 10a, 14b and (b) cumulus legs-7b-e, 11c, 15d at 150 m (black) and in-cloud (green) levels. The rain-rate cumulative frequency at a given drop size gives the fraction of rain rate obtained from all the drops equal to or smaller than that drop size.](image1)

![Fig. 12. Drop-size distributions (dN/dD) in m^{-3} mm^{-1} for (a) stratocumulus modules (6a-b, 10a, 14b) and (b) cumulus modules (7b-e, 11c, 15d) at 150 m (black line) and in-cloud (green line) levels. Exponential (blue) and lognormal (red) fit lines are plotted for each 150 m (solid) and in-cloud (dotted) RSD. The mean diameter $D$ of the exponential fit lines, along with the geometric median diameter $D_g$ and geometric spread $\sigma$ of the lognormal fit lines are denoted in the plots, color-coded as green for in-cloud and black for 150-m legs.](image2)
cycle is expected (e.g., Eastman and Wood 2016), in that shortwave absorption can decouple the boundary layer during day. Nevertheless, the advecting and deepening cloudy air mass can still recouple to the surface on the subsequent night. The pace of cloud transitions reveals a further connection to the boundary layer deepening, summarized in Fig. 14. The slower transitions (>36 h) occur on the third day, beginning between 0500 and 0900 local time. This describes the easternmost two transitions of 17 July, beginning from the RF6a air mass. Their advection keeps them further east, where the boundary layer deepens more slowly. One of the trajectories emanating from RF10ab on 27 July is similar. We speculate that these boundary layers are still able to flux moisture from the surface to the cloud layer during the second night, but that precipitation-induced decoupling during the second night preconditioned the air mass for morning breakup. The other transitions originating on 17 July begin on the late afternoon/early evening of the first day, between 1600 and 2000 local time. These air masses deepened more over the course of 2 days. A consistent interpretation is that the boundary layer was simply too deep by the end of the daytime hours to support even an intermittent coupling to the surface during the subsequent night.

Statistically, precipitation is known to be closely linked to the boundary layer depth, shown, for example, in O et al. (2018) for all of the SCT regions. Precipitation and boundary layer depth are further related to the timing of the transitions in Figs. 14b and 14c. The boundary layers that deepen the most also transition more quickly (Fig. 14b), and the swiftest transitions are accompanied by the largest changes in mean rain rates (Fig. 14c). Faster transitions can originate from deeper stratocumulus layers with less precipitation reaching the surface (e.g., 7 August, module 14b), or from shallower stratocumulus with more precipitation reaching the surface (17 July). The slower transitions show similar relationships between boundary layer depth and rain rates, with the hours to transition primarily set by the timing of the diurnal cycle. These cases do make clear that the rain rates adjust quickly to the evolving boundary layer depth, with time lags that must be <1 day, consistent with Eastman et al. (2016).

7. Change in cloud base, near-surface relative humidity, and subcloud evaporation

A lower relative humidity (RH) near the surface will correspond to a higher lifting condensation level, with a higher cloud base height. Both a higher cloud base, and a drier subcloud atmosphere, will encourage more evaporation of larger drops. These considerations can be examined with the in situ near-surface relative humidity. The relative humidity averaged over both the entire and nonraining portions of the 150-m subcloud legs and the lidar-derived cloud base heights are indicated as a function of longitude for the Lagrangian trajectory beginning and endpoints (Fig. 15). The near-surface RH varies within the stratocumulus region, from 85% to 90% on 17 July (RF6a-b), corresponding to cloud bases near 400 m, to 70%–80% for 27 July and 7 August, corresponding to cloud bases between 700 and 950 m. A drop with a diameter of 180 μm is capable of reaching the surface, at a distance of 400 m. In contrast, the mean cloud base heights of approximately 900 m on 27 July and 7 August (RF10b and RF14b) require a drop diameter of at least 400 μm before the surface can be reached, for a near-surface relative humidity of 75% increasing with height. The near-surface RH and cloud bases are more similar within the cumuli regions, ranging between 75%–85% and 700–1000 m, respectively. This corresponds to a decrease in near-surface RH for the 17–19 July transition, while the near-surface RH increases in the two other cases. Zhou et al. (2015), using ship-based data from further south of the CSET stratocumulus modules, linked a lower near-surface relative
humidity and higher LCLs to increased cloud-top entrainment throughout the SCT. The increase in the near-surface RH for the 27–29 July and 7–9 August transitions shown here, may instead reflect a stronger temperature and moisture stratification as the boundary layer decouples, similar to cumulus-under-stratocumulus conditions in the southeast Atlantic (Zhang and Zuidema 2019). Drying of the subcloud layer does not always occur throughout the SCT, and indeed as precipitation becomes more intense, their cold pools and residual moist patches can become important for organizing further convection (Li et al. 2014; Bretherton and Blossey 2017).

The change in the near-surface relative humidity as the air masses advect equatorward will affect the subcloud evaporation profile, with drier conditions encouraging more evaporation. This is assessed using both the subcloud evaporation model applied to all three flight pairs, and the radar-lidar retrievals for the rain characteristics for July 17–19 transition. As will be shown these estimates indicate that the subcloud latent cooling typically increases across the transition, although likely the least for July 17–19 transition because the initial subcloud evaporation is large.

a. Subcloud evaporation model

Subcloud profiles of the rainwater content are calculated using the one-dimensional model described in section 2d, initialized with the in-cloud raindrop size distributions for all the modules. Examples are shown for RF6a and RF7c in Fig. 16. These are used to estimate the subcloud evaporation between cloud base and surface, with the latent cooling reported in Table 3 for each module. As expected (e.g., Wood 2005b),
much (though not all) of the stratocumulus precipitation evaporates close to the cloud base, where the latent cooling can support the buoyancy of moist near-surface parcels further, supporting the cloud layer. In the example shown, the vertically integrated subcloud evaporation increases for RF7c, while much more of the rain is also able to reach the surface. This will lower the evaporative cooling profile, all else equal, supporting further surface fluxes. RF7c includes some of the most pronounced precipitation sampled during CSET, with the model indicating increased subcloud evaporation even as the drops shift to larger sizes. The same holds true when the evaporative fluxes are averaged together for all the stratocumulus and cumulus modules. Overall, in the mean, the model values support the idea that the subcloud evaporation increases throughout the transition regardless of the initial environmental subcloud relative humidity.

b. Radar-lidar retrievals

The subcloud evaporation is also estimated from the difference in the radar-lidar retrievals of rainwater content at cloud base and near the surface (Schwartz et al. 2019) for RF6a and RF7c; the remote sensing measurement were made during the subcloud legs of the modules, offset by 10 min from the in-cloud legs used to initialize the evaporation model. Rain shafts were well sampled by the radar and lidar during the two subcloud legs (Fig. 17, top row). Retrievals of the other rain properties are also shown, to help understand differences from in situ values. The retrieved rain rates, rainwater content, rain effective diameter and raindrop number concentrations are most pronounced for the

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**Fig. 16.** Subcloud rainwater content (RWC, g m\(^{-3}\)) profiles calculated from the evaporation model and initialized with the in-cloud rainwater content, for module (left) 6a and (right) 7c. The cloud-base height (H), subcloud evaporation flux (F), in-cloud precipitation frequency (Freq), and relative humidity at the nearest 150-m level leg from 1-s samples lacking liquid (RH\(_{150}\)) are shown in the title of each plot. The subcloud evaporation flux is calculated as indicated in section 2. The mean GV altitude during the in-cloud levels are also shown within the figures.

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**Table 3.** HCR leg mean cloud base height, CBH (m); frequency of raining samples at in-cloud legs; leg mean relative humidity at 150 m, RH\(_{150}\) (%); in-cloud leg mean CDP concentration, N\(_{\text{CDP, in-cloud}}\) (cm\(^{-3}\)); in-cloud leg mean 2DC concentration, N\(_{\text{2DC, in-cloud}}\) (L\(^{-1}\)); leg-mean RWC at 150 m, RWC\(_{150}\) (g m\(^{-3}\)); leg mean RWC at in-cloud level, RWC\(_{\text{in-cloud}}\) (g m\(^{-3}\)); leg mean rain rate at 150 m, R\(_{150}\) (mm h\(^{-1}\)); leg mean rain rate at in-cloud level, R\(_{\text{in-cloud}}\) (mm h\(^{-1}\)); and leg mean evaporative flux between cloud base and 150-m altitude, F (W m\(^{-2}\)), calculated using the subcloud evaporation model discussed in section 2c.

<table>
<thead>
<tr>
<th>Stratocumulus</th>
<th>Cumulus</th>
</tr>
</thead>
<tbody>
<tr>
<td>6a</td>
<td>6b</td>
</tr>
<tr>
<td>CBH (m)</td>
<td>422</td>
</tr>
<tr>
<td>Frequency</td>
<td>0.6</td>
</tr>
<tr>
<td>RH(_{150}) (%)</td>
<td>88</td>
</tr>
<tr>
<td>N(_{\text{CDP, in-cloud}}) (cm(^{-3}))</td>
<td>14</td>
</tr>
<tr>
<td>N(_{\text{2DC, in-cloud}}) (L(^{-1}))</td>
<td>24</td>
</tr>
<tr>
<td>RWC(_{150}) (g m(^{-3}))</td>
<td>0.05</td>
</tr>
<tr>
<td>RWC(_{\text{in-cloud}}) (g m(^{-3}))</td>
<td>0.041</td>
</tr>
<tr>
<td>R(_{150}) (mm h(^{-1}))</td>
<td>0.54</td>
</tr>
<tr>
<td>R(_{\text{in-cloud}}) (mm h(^{-1}))</td>
<td>0.29</td>
</tr>
<tr>
<td>F (W m(^{-2}))</td>
<td>45</td>
</tr>
</tbody>
</table>
precipitation shaft within RF7c, with rain rates at 143.4°W reaching 3 mm h\(^{-1}\)—comparable if slightly less than the observed in situ rain rates of 3–9 mm h\(^{-1}\) (Fig. 10). Notable is the lack of retrievals for the stratocumulus cloud.

The remote sensing retrievals and in situ datasets are compared more clearly within Fig. 18. The lowest altitude at which retrievals are available is 360 m for RF6a, and 380 m for RF7c, whereas the in situ values are slightly lower, at 150-m altitude. The retrieved rain rates are less, by one order of magnitude, than the rain rates deduced from the in situ probes, based on rain frequencies that are opposite to those observed—too few values are retrieved for the stratocumulus module. In contrast, the retrieved rainwater contents tend to exceed those derived from the 2DC probes. The retrieved raindrop number concentrations are underestimated, while the retrieved effective diameters are slightly overestimated. The neglect of the lighter stratocumulus precipitation is because of a retrieval requirement that reflectivity-to-backscatter ratios exceed 30 dB. This is intended to exclude background aerosol, but can also exclude light drizzle, for which the lidar echo becomes
comparable with that of background aerosols. We hypothesize that Mie scattering, not fully accounted for within the retrieval, may explain the underestimate of the heavier precipitation.

Despite the problems, we can still use the retrieved estimates of the rain rates to gauge if the subcloud evaporation increases during the SCT. It does—from 0.6 W m\(^{-2}\) during module 6a to 2.4 W m\(^{-2}\) within module 7c, calculated from the difference between the rain rate at cloud base and the lowest retrieval altitude and converted to a latent heating. This is substantially less—by more than an order of magnitude—than the estimate provided by the evaporation model but does indicate an increase in the subcloud latent cooling across the transition, consistent with the evaporation model. The moistening and cooling will help compensate for the increasing entrainment of warm, dry air from above the inversion.

8. Summary and conclusions

Few in situ observations have been available to date on the precipitation characteristics of SCTs. Three flight pairs, on 17–19 July, 27–29 July, and 7–9 August 2015 sampled cloud transitions that clearly began with stratocumulus and allow for satellite views unimpeded by high clouds. The stratocumulus clouds were sampled near and to the west of 130°W, 40°N, to the north-northwest of the region with the climatological maximum in stratocumulus (20°–30°N, 120°–130°W). The stratocumulus clouds sampled on the three different days include a low-lying (cloud base at 400 m), pristine
(\(N_d = 22 \, \text{cm}^{-3}\)) cloud with precipitation reaching the surface on 17 July, a higher-altitude (cloud base of 680 m) polluted (\(N_d = 225 \, \text{cm}^{-3}\)) cloud with less precipitation reaching the surface on 27 July, and a higher-altitude thinner, pristine (\(N_d = 22 \, \text{cm}^{-3}\)) cloud with almost no precipitation reaching the surface on 7 August. The air masses thereafter evolve 2 days later to cumuli reaching 2 km, over waters that are 4–5 K warmer. Precipitation cleanses the mid-July boundary layer, resulting in ultraclean cloud (\(N_d < 1 \, \text{cm}^{-3}\)) documented more thoroughly in Wood et al. (2018) and Kuan-Ting et al. (2018). This does not occur for the two other cases. Of note are the high near-surface precipitation rates on 17 July (leg \(2\)) documented more thoroughly in Wood et al. (2018) and Kuan-Ting et al. (2018). Such precipitation rates will remove the boundary layer cloud condensation nuclei through coalescence alone within approximately five hours (Wood 2006), and with little replacement from the free troposphere or surface, easily explain the transition to the ultraclean conditions.

An evaluation of the subcloud evolution in the relative humidity and corresponding cloud bases found that the subcloud layer could both dry and moisten, depending on the flight pair. A well-coupled low-lying stratocumulus cloud evolved into a drier subcloud layer, indicating the influence of cloud-top entrainment. In contrast, the deeper, stratocumulus cloud layers with drier subcloud layers moisten throughout the transition, through either rain evaporation and/or a lack of upward ventilation of the surface fluxes as the deeper boundary layer decouples. The latter process in particular will influence the ability of shallow convection to further organize (e.g., Li et al. 2014; Bretherton and Blossey 2017).

A GOES-obtained cloud fraction analysis indicates that transitions of stratocumulus to cumulus with cloud fraction below 50%, can begin either in the morning or late afternoon during sunlit hours, suggesting that nighttime recoupling of the boundary layer, or the lack thereof, is important for setting the overall time scale. The boundary layers that deepen the fastest, typically have higher rain rates at the end of transition and also transition more quickly. In contrast to conclusions drawn from space-based larger sample sizes (Eastman and Wood 2016), the stratocumulus deck with the strongest coupling to the surface on 17 July, is not clearly more persistent than one on 7 August that occupies a deeper boundary layer and is otherwise similar.

A further in-depth analysis will be required to examine how the radar-lidar retrievals can be improved for the clouds sampled during CSET. We speculate that the presence of raindrops with diameters greater than 1 mm may affect retrievals of rain rates from the HIAPER Cloud Radar, which has a wavelength of 3.22 mm, due to Mie scattering. If so, this will contribute to an underestimate of the retrieved rain rates. While occasional, the large drop sizes contribute disproportionately to the overall rain rate, with the in situ drop size distributions clearly evolving to larger sizes throughout the SCT, consistent with collision-coalescence within deeper clouds (e.g., Hudson et al. 2015). The influence of Mie scattering will be pursued more rigorously in an upcoming paper.

Arguments can be made for both microphysical and thermodynamic contributions by precipitation to the pace of the transition. As the raindrop size distributions shift toward larger drops that are more capable of reaching the surface and leaving the atmosphere, the removal of aerosol becomes reinforced (e.g., Yamaguchi et al. 2017). The subcloud evaporation is found to increase throughout the SCT, in both a simple model and based on the radar/lidar retrieval. While each estimate may be flawed, the two independently derived values reinforce each other in sign. The increased evaporation, combined with the shift to larger drop sizes will place more of the evaporation near the surface, encouraging more surface-driven cumulus cloud development, as the SCT progresses. The coassociation between boundary layer depth and precipitation within this limited sample size prevents a comprehensive assessment of the precipitation influences, but the variety in the three individual cases lends them well to constraining more detailed future modeling studies.

Acknowledgments. MS, PZ, and BA acknowledge support from NSF Grant AGS-1445832. VG would like to acknowledge National Science Foundation (NSF) Grant AGS-1445831 597 awarded to the University of Chicago and the U.S. Department of Energy’s (DOE) Atmospheric 598 System Research (ASR). NASA Langley provided the GOES-15 cloud-top temperature retrievals. The merged, quality-controlled radar/lidar 2Hz data-set placed on a uniform georeferenced grid is available at https://doi.org/10.5065/D63T9FM0 and their retrievals are available from https://anl.app.box.com/s/g4tugv86ddzfhihsp47fcqy19kkk9aze and from VG. All other datasets are available through the NCAR EOL field catalog at http://catalog.eol.ucar.edu/cset. We thank three anonymous reviewers for their thoughtful comments.

APPENDIX

Raindrop Size Distribution Fitting

The accuracy of truncated lognormal and exponential fits is assessed for the 2DC raindrop size distributions

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(RSDs) shown in Fig. 11 for the in-cloud and 150-m legs. The probability density function (pdf) for exponential fitting truncated at 75 μm is
\[
N(D)dD = \frac{1}{D - D_o} \exp \left( -\frac{D - D_o}{D - D_o} \right),
\]
where \(D_o\) is 75 μm, \(D\) is the median of 2DC bin diameters, and \(D\) is the mean diameter of drops >75 μm (Comstock et al. 2004, Wood 2005b); \(D\) is shown for both in-cloud and 150-m legs in Fig. 11. The pdf for the lognormal fitting is given by
\[
N(D)dD = \frac{1}{D\sqrt{2\pi\sigma^2}} \exp \left( -\frac{(\ln D - \mu)^2}{2\sigma^2} \right),
\]
where \(\mu\) and \(\sigma\) are the mean and standard deviations of the lognormally transformed variables, respectively (Cho et al. 2004). The geometric mean diameter \(D_g(=e^\mu)\) represents the median of the size distribution, while \(\sigma\) is the geometrical standard deviation or the width of the distribution (vanZanten et al. 2005). The analytical relationships in Feingold and Levin (1986) extend the truncation of the lognormal pdf at the boundaries beyond 75 μm and 3.2 mm; \(D_g\) and \(\sigma\) are also shown for all the in-cloud and 150-m legs along with \(D\) in Fig. 11. Since the RSD is skewed toward the small drop sizes, \(D\) (mean) is larger than \(D_g\) (median) in most cases. A visual comparison of the fit lines for all the cases indicates the lognormal fitting provides a better representation of the RSD at large drop sizes, although for drops smaller than 0.15 mm in the in-cloud legs, the exponential distribution may be better. The shift of the RSD toward larger drop sizes with the transition can be seen from the increase in \(D_g\) from 0.1 mm within all stratocumulus clouds to 0.2–0.5 mm in cumulus clouds. The value \(\sigma\) also increases along the transition and toward the surface.

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