Lagrangian Evolution of the Northeast Pacific Marine Boundary Layer Structure and Cloud during CSET^①

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ABSTRACT

Flight data from the Cloud System Evolution over the Trades (CSET) campaign over the Pacific stratocumulus-to-cumulus transition are organized into 18 Lagrangian cases suitable for study and future modeling, made possible by the use of a track-and-resample flight strategy. Analysis of these cases shows that 2-day Lagrangian coherence of long-lived species (CO and O₃) is high ($r = 0.93$ and 0.73 , respectively), but that of subcloud aerosol, MBL depth, and cloud properties is limited. Although they span a wide range in meteorological conditions, most sampled air masses show a clear transition when considering 2-day changes in cloudiness (-31% averaged over all cases), MBL depth ($+560$ m), estimated inversion strength (EIS; -2.2 K), and decoupling, agreeing with previous satellite studies and theory. Changes in precipitation and droplet number were less consistent. The aircraft-based analysis is augmented by geostationary satellite retrievals and reanalysis data along Lagrangian trajectories between aircraft sampling times, documenting the evolution of cloud fraction, cloud droplet number concentration, EIS, and MBL depth. An expanded trajectory set spanning the summer of 2015 is used to show that the CSET-sampled air masses were representative of the season, with respect to EIS and cloud fraction. Two Lagrangian case studies attractive for future modeling are presented with aircraft and satellite data. The first features a clear Sc–Cu transition involving MBL deepening and decoupling with decreasing cloud fraction, and the second undergoes a much slower cloud evolution despite a greater initial depth and decoupling state. Potential causes for the differences in evolution are explored, including free-tropospheric humidity, subsidence, surface fluxes, and microphysics.

1. Introduction

Marine boundary layer (MBL) clouds are important contributors to Earth's energy budget, due to their extensive spatial coverage, high albedo, and relatively warm cloud-top temperature (Hartmann et al. 1992).

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As a result, these clouds have long been subject to observational and modeling scrutiny (Albrecht et al. 1988; Stevens et al. 2005; Wood, 2012). An important climatological feature is the subtropical stratocumulus (Sc) to trade cumulus (Cu) transition (SCT), where shallow Sc-topped MBLs over cool coastal waters advect westward and equatorward in the trade winds, the boundary layer deepens and eventually the Sc cloud breaks up, leaving behind a cumulus MBL with lesser cloud fraction.

The dynamics of the marine cloud-topped boundary layer has a rich history of investigation. Early seminal work was performed by Lilly (1968), who explored

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cloud-topped MBL mixing, crucially describing the role of cloud-top radiative cooling in driving MBL mixing (in contrast to continental boundary layers predominantly driven by surface fluxes). [Randall \(1980\)](#) examined the role of cloud-top entrainment in generating additional instability and Sc breakup. [Nicholls \(1984\)](#) and [Nicholls and Leighton \(1986\)](#) explored in depth the decoupling of the cloud versus subcloud layers in marine stratocumulus, noting the role of cumulus clouds in recoupling the two layers. From a climatological perspective, [Klein and Hartmann \(1993\)](#) found that lower-tropospheric stability (LTS) correlated strongly with MBL cloudiness and highlighted the importance of a strong inversion for a cloudy MBL. The 1992 ASTEX field campaign provided a great deal of insight into the dynamics of the transition ([Bretherton and Pincus 1995](#); [Bretherton et al. 1995a,b](#)), by noting the link between decoupling and breakup in two Lagrangian cases, and suggesting a possible microphysical role for aerosol in the transition in some cases ([Albrecht et al. 1995](#)). [Krueger et al. \(1995\)](#) reproduced the transition using numerical simulations forced only by increasing sea surface temperature (SST). [Wyant et al. \(1997\)](#) and [Bretherton and Wyant \(1997\)](#) developed a comprehensive model for the SCT, in which boundary layers advecting over warmer waters and away from subtropical maxima in subsidence deepen and decouple, developing a cumulus layer underneath the overlying stratocumulus. The cumulus further dries out the overlying stratocumulus by entraining free-tropospheric air through a weaker inversion, eventually leading to a cumulus-dominated MBL. This model (hereafter W&B model) also helps explain the LTS-cloudiness correlation found by [Klein and Hartmann \(1993\)](#). [Wood and Bretherton \(2006\)](#) define the estimated inversion strength (EIS), a refinement of LTS applicable over a broader range of temperatures.

Further studies of the relationship between MBL vertical structure and cloud evolution have also added to our understanding of the SCT. Consistent with the W&B model, [Wood and Bretherton \(2004\)](#) showed that satellite estimates of decoupling and boundary layer depth are well correlated. The free-tropospheric relative humidity (FT RH) affects low cloud cover by altering both the net longwave radiative cooling and the entrainment at cloud top ([Mauger and Norris 2010](#); [Kawai et al. 2017](#); [Adebiyi and Zuidema 2018](#)), and the magnitude of the effect of FT RH anomalies on cloud cover is sensitive to the vertical distribution of moisture ([Eastman and Wood 2018](#)). Large-eddy simulation (LES) studies focusing on cloud-top entrainment (e.g., [Lock 2009](#); [Xiao et al. 2011](#)) suggest that Sc breakup is affected by the cross-inversion moisture jump as well as the temperature jump.

Observationally inferred relationships are dependent on the time scale considered for MBL evolution.

[Eastman et al. \(2017\)](#) found that on seasonal time scales LTS and MBL depth are uniformly anticorrelated, consistent with W&B, but regions of positive correlation between LTS and depth emerge when the seasonal-scale variability is removed, possibly due to other drivers such as FT RH. That work and others have also highlighted the utility of a Lagrangian perspective for studying drivers of MBL evolution, as there is a lag in the MBL response to changing stability ([Klein et al. 1995](#)) and upwind factors must be controlled for ([Mauger and Norris 2010](#)). However, [Eastman et al. \(2016\)](#) showed that many MBL cloud properties decorrelate faster (in a day or less) following advecting boundary layer air columns (Lagrangian analysis) than in a fixed Eulerian perspective; it was postulated that this was due to the advection of the MBL across persistent but spatially variable cloud-controlling factors.

MBL aerosol concentration affects clouds, primarily by modulating the cloud droplet number concentration N_d ([Twomey 1974](#)). Through microphysical feedbacks, this will also influence cloud macrophysical evolution, sometimes called the “lifetime” effect ([Albrecht 1989](#)). The lifetime effect involves many small-scale cloud processes and depends on the cloud regime, so it is a major source of uncertainty in modeling anthropogenic aerosol effects on climate. Process modeling studies based on LES have identified two competing mechanisms, working through entrainment and precipitation, for liquid-phase MBL cloud lifetime effects. [Ackerman et al. \(2004\)](#), [Jiang et al. \(2006\)](#), [Xue and Feingold \(2006\)](#), [Bretherton et al. \(2007\)](#), [Wood \(2007\)](#) and others showed that higher N_d , through its impact on cloud droplet size, increases evaporation–entrainment feedbacks that reduce cloud liquid water content. These studies also noted that lower N_d accelerates the onset of precipitation in thicker clouds, which can decouple a Sc-capped MBL ([Stevens et al. 1998](#)) and may help transition Sc to Cu ([Sandu and Stevens 2011](#); [Berner et al. 2013](#); [Yamaguchi et al. 2017](#)). Both mechanisms are challenging to represent in global climate model parameterizations because of their dependence on interacting small-scale processes.

Observational constraints are required to evaluate modeling of aerosol–cloud–precipitation interactions, and numerous past studies have used in situ and remote sensing studies to this end. For instance, [Bretherton and Pincus \(1995\)](#) documented two ASTEX Lagrangian case studies. The more polluted air mass retained a persistent Cu under Sc structure despite being decoupled, while the cleaner one transitioned from stratocumulus to cumulus. However, observationally assessing aerosol effects on cloud evolution is challenging because meteorological differences between cases also strongly impact

cloud formation. The aim of the Cloud System Evolution over the Trades (CSET) campaign was to provide a comprehensive in situ dataset that could be used to better assess aerosol–cloud interaction and our ability to model it, with Lagrangian sampling for many cases supporting an improved separation of meteorological and aerosol controls on MBL clouds.

The goal of this paper is to use this new set of aircraft observations to better understand and document cloud and boundary layer controlling factors in the SCT and provide some attractive case studies for detailed process modeling. The data were collected during the CSET field campaign (Albrecht et al. 2019), which took place 1 July–15 August 2015 in the northeast Pacific subtropical stratocumulus region. The primary observational platform for CSET was the NSF/NCAR Gulfstream V (GV), flying between Sacramento, California (CA), and Kona, Hawaii (HI). Low-level sampling was carried out throughout the depth of the MBL and lower free troposphere using a suite of in situ and remote sensing instrumentation. CSET employed a Lagrangian approach pairing outbound (CA–HI) flights to return (HI–CA) flights two days later. As we use these terms throughout the paper, note that in the context of Lagrangian evolution, the “outbound” data corresponds to the initial upwind sampling, and the “return” data corresponds to the downwind sampling 2 days later. MBL air masses sampled on the outbound flight were tracked using 500-m constant-height HYSPLIT trajectories, and then resampled again on the return flight. Many of these air masses transitioned from stratocumulus-capped on the outbound flight to primarily shallow cumulus on the resampling portions of the return flight, and a diversity of aerosol conditions were encountered. For an in-depth description of the campaign and instrumentation, see Albrecht et al. (2019).

In this paper, we present a Lagrangian view of the subtropical MBL evolution of CSET sampled/resampled air masses, combining aircraft, geostationary satellite, and reanalysis data. Our work complements an Eulerian analysis of the CSET observations presented in Bretherton et al. (2019), which uses compositing techniques to take advantage of the generally similar sampling during all the flight pairs and develop a statistical representation of the SCT. Bretherton et al. (2019) analyzes the mean state and variability of clouds, aerosols and precipitation along a nominal transect between California and Hawaii. Mean states for these quantities are used to assess the capabilities of reanalysis and reanalysis-nudged global climate model single-column output for this region. Bretherton et al. (2019) also include a comparison between satellite retrieved N_d and in situ sampling.

FIG. 1. Example of one of seven CSET mission pairs, here showing RF08/RF09, flown on 22 Jul and 24 Jul 2015, respectively. Ferry legs are shown in black, while low-level sequences are the blue-and-orange portions of each flight. Tracking trajectories initialized on the outbound flight path are shown in red, with a dot marking every 6 h of advection. (inset) CSET flight sequence, showing progression through below-cloud, in-cloud, porpoise, and above-cloud legs relative to cloud base and cloud top.

We first describe how the Lagrangian cases are constructed (section 2a), additional datasets (sections 2b and 2c), and how MBL variables are estimated (section 2d). A statistical analysis of overall MBL Lagrangian evolution from all CSET cases and from the expanded trajectory set are presented in section 3a, while two contrasting case studies recommended for future modeling are presented in section 3b. A discussion of results and conclusions follow in section 4.

2. CSET measurements and data analysis methods

a. Lagrangian case data and methodology

1) CSET FLIGHTS AND DATA

As described by Albrecht et al. (2019), the CSET flights consisted of a ferry leg (i.e., high-altitude transit flight) to reach an area of interest in the remote MBL, followed by approximately 1500 km of low-level sampling, and then another ferry leg to complete the transit to either Kona (outbound flights) or Sacramento (return flights). The low-level sampling consisted of a series of 3–5 sequences of a fixed set of flight maneuvers per flight (Fig. 1, inset): descent to 150 m above the sea surface for a 10-min “below-cloud” level flight leg, ascent to 10-min “in-cloud” level leg just above cloud base, 10-min “porpoise leg” (i.e., repeating sawtooth) through

cloud top and inversion layer, and ascent to 1500 m above cloud top for 10-min “above-cloud” level leg, then descending again to begin the next sequence. Each sequence lasted approximately 50 min and sequences were lettered alphabetically, with A being the first sequence of each flight, B the second, etc. An example of a CSET flight pair is shown in Fig. 1. This figure shows the outbound flight RF08 from California to Hawaii, tracking trajectories initialized on the RF08 flight path, and subsequent RF09 return flight intercepting the trajectories (for CSET, even-numbered flights were outbound CA-HI, and odd-numbered flights were return HI-CA). Distortion of the sampled line due to regions of convergence and divergence resulted in a dog-leg (kink in flight path) in RF09. The flight plans were chosen so that the return flight resampled 500 m constant-level trajectories from the outbound flight, calculated as discussed in section 2b.

GV in situ instrumentation (Albrecht et al. 2019) included a standard set of temperature and humidity probes, CO and O₃ measurements, an Ultrahigh Sensitivity Aerosol Spectrometer (UHSAS) measuring size-resolved aerosol from 60 to 1000 nm diameter, a Cloud Droplet Probe (CDP, measuring 2–50 μ m diameter drops), precipitation probes and broadband radiometers. For remote sensing, the plane was equipped with the HIAPER Cloud Radar (HCR) and a High Spectral Resolution lidar (HSRL), both of which could be aimed up or down, as well as broadband radiometers and dropsondes.

2) CONSTRUCTION OF CSET LAGRANGIAN CASES

In the context of the CSET campaign, “Lagrangian” refers to following the MBL air masses that were tracked by trajectories and twice sampled by the aircraft. To organize the flight and trajectory data into 18 Lagrangian cases (2–3 per flight pair), the tracking trajectories were used to identify that portions of the outbound flight were resampled by which portions of the return flight. An additional constraint was that each case had to contain at least one full sampling sequence (below-cloud leg through above-cloud leg) on both the outbound/upwind and return/downwind flight. Due to variability in divergence of the sampled air mass, either the outbound or return aircraft data could span a longer distance. The resulting cases represented air masses on the scale of 500 km and spanned a significant amount of meso-scale variability. Cases are numbered in chronological order and are referenced as L01–L18 (for Lagrangian cases 1–18). Collocated satellite measurements and reanalysis data along the flight paths and the connecting trajectories complement the aircraft observations.

FIG. 2. Lagrangian case studies (a) L06 (from RF06/RF07) and (b) L10 (from RF10/RF11) showing the location of the flight portions (thicker blue/orange lines) and air mass trajectories (thinner lines). Insets show enlarged GOES cloud field (from a $2^\circ \times 2^\circ$ box centered on trajectories 2.3 and 6.0, respectively) at the time of initial sampling, a day later, and two days later at the time of re-sampling. Hour 49 is used instead of 48 due to partially missing data. Pink circles indicate sequences of flight data.

Two such cases are shown in Fig. 2. Maps for other cases can be found in the supplemental figures in the online supplemental material, along with a table showing key BL variables (cloud fraction, N_d , stability, etc.) for every case. As an example, the case in Fig. 2a, L06, consists of aircraft data from the middle of RF06 (sequences B and C, thick blue line), which was resampled at the beginning of RF07 (sequence A, thick orange line). These two flight portions were connected by 5 tracking trajectories (T1.6–T2.6, thinner colored lines). “For both outbound and return flights, aircraft data are averaged over the appropriate legs (e.g. below-cloud or in-cloud) from all relevant sequences (e.g. RF06 BC, RF07 A).” For instance, the Lagrangian change in droplet number concentration (N_d) for L06 is the difference between the initial aircraft N_d from RF06, sequences B and C,

in-cloud legs, and the final aircraft N_d from RF07, sequence, A, in-cloud leg. Trajectory data are averaged over all trajectories in the case except where explicitly noted. For instance, the mean subsidence experienced by L06 is the 2-day time-mean ERA5 subsidence along all 5 trajectories T1.6–T2.6 from RF06/07, then again averaged over trajectories (see [section 2d](#) for more detail on subsidence and entrainment).

To support future process modeling studies and to highlight the range of MBL cloud evolution sampled in CSET, we present in detail two Lagrangian case studies shown in [Fig. 2](#): L06 in [Fig. 2a](#), and L10 in [Fig. 2b](#), (corresponding to the easternmost air mass from RF10, sequence A, which was resampled on RF11 DE). Both cases start out cloudy and experience a strong decrease in EIS, but while the L06 case experiences a classic SCT, the L10 case remains cloudier. The two cases differ in initial MBL depth, decoupling, microphysics, and large-scale forcings. Note that the agreement between outbound flight number and case number (L06 coming from RF06, L10 from RF10) is coincidental.

b. Satellite and reanalysis data

In addition to the GV measurements, satellite and reanalysis data are used. Where possible, these are compared to aircraft observations for additional validation. The satellite data consists of cloud property fields generated from *Geostationary Operational Environmental Satellite-15 (GOES-15)* observations using algorithms originally developed for the NASA Clouds and the Earth's Radiant Energy System (CERES) project ([Minnis et al. 2008a, 2011](#)) and adapted for application to other imagers on geostationary satellites ([Minnis et al. 2008b](#)). Except where noted, all retrievals are averaged over a $2^\circ \times 2^\circ$ box centered on the coordinates in question (e.g., aircraft or trajectory position) and are at an hourly resolution. The 2° box was empirically chosen to balance the competing interests in reducing noise in box-averaged quantities while avoiding including observations from regions subject to significantly different large-scale forcings; a comparison of the GOES cloud fraction estimate to that derived from a radar-lidar cloud mask can be found in [Bretherton et al. \(2019\)](#).

Supplemental data are drawn from the European Centre for Medium-Range Weather Forecasts (ECMWF) Reanalysis, version 5 (ERA5) [[Copernicus Climate Change Service \(C3S\) 2017](#)], for large-scale atmospheric conditions. Trajectories are generated using the Hybrid Single-Particle Lagrangian Integrated Trajectory model (HYSPLIT; [Stein et al. 2015](#)), using the National Centers for Environmental Prediction (NCEP) Global Forecast System (GFS) forecasts and the Global Data Assimilation System (GDAS) analyses as provided by the National

Oceanic and Atmospheric Administration Air Resources Laboratory (NOAA ARL) archive.

c. Expanded trajectory set

During CSET, 53 trajectories within the 18 cases were used in total to track sampled air masses. As the satellite and reanalysis data are not restricted to the context of the CSET flights, we generated a larger dataset of approximately 1800 trajectories, spanning July–August 2015. This is used in [section 3a](#) to validate the representativeness of the CSET sampled air masses and provides a large sample size for testing certain hypotheses that do not rely on any aircraft data. The trajectories were initialized at 0000 UTC for every day in the two-month span, on the same grid used for CSET flight planning. The resulting trajectories were filtered for direction to exclude 800 trajectories going toward land or the extratropics. Satellite and reanalysis data were extracted along the remaining 1000 trajectories.

d. Calculation of MBL variables

For aircraft data, some variables are drawn only from certain legs (such as below-cloud) in each sequence and some are only from cloudy observations (defined as $RH > 95\%$ and CDP liquid water content $q_l > 0.01 \text{ g m}^{-3}$) or conversely noncloudy observations. Aircraft cloud droplet number concentration (N_d) is measured by the CDP and averaged only over the cloudy time steps on the in-cloud leg. Accumulation-mode aerosol concentration (all particles with diameters between 100 and 1000 nm) is measured by the UHSAS and averaged over the below-cloud leg. Aircraft SST estimates are averages of the radiometric surface temperature from the below-cloud leg. Carbon monoxide (CO) and ozone (O_3) measurements were averaged over the below-cloud leg as well. Note that a particular case may span 300–600 km of sampling, especially on the return flight in which it may encompass multiple sequences. Across this distance, significant mesoscale variability is inevitably encountered, and is only partly smoothed by leg averaging. This adds unavoidable noise to relationships between cloud and environmental conditions.

Cloud fraction is defined as the fraction of GOES pixels in a $2^\circ \times 2^\circ$ box (centered on aircraft location or trajectory) that have successful retrievals of cloud type that are not clear sky. We additionally only focus on warm low liquid cloud fraction by selecting for cloud tops under 4 km with liquid tops. A random overlap assumption is made for pixels masked by deep/high clouds. Cloud droplet number concentration is calculated from GOES retrievals following [Painemal and Zuidema \(2011\)](#) using cloud optical thickness and cloud effective radius. Both CF and N_d are averaged along all

outbound and return sequences from that Lagrangian case, and for N_d , the 75th percentile of values for warm low liquid cloud pixels is used to mitigate for possible biases caused by subpixel cloud inhomogeneity. Cloud fraction, as well as cloud-top height, span the full diurnal cycle, whereas N_d and liquid water path retrievals are daytime only.

Inversion base height z_i (which typically is very close to the top of any extensive MBL stratocumulus), decoupling, and cross-inversion temperature and moisture jumps are calculated from the aircraft profiles, down from either the above-cloud or ferry leg to the below-cloud leg. As such, the number of profiles in each case (outbound or return) is equal to the number of sequences (e.g., case L10 has one profile from the outbound flight and two from the return flight). The estimation of the inversion base height z_i , inversion layer, jumps in q and T across the inversion, and decoupling are all related. The [appendix](#) describes in detail how these quantities are measured from aircraft profiles and reanalysis. In brief, the boundary layer inversion middle is identified as the level of greatest lapse rate (after smoothing, forced between 300 and 3000 m). The inversion base height, z_i , is the nearest level below the center where the lapse rate is (nearly) moist adiabatic, and the inversion top height is the nearest level above the center where the same is true. This method was found to work well with both aircraft profiles and reanalysis, with only a small bias in the inversion center height (see the [appendix](#) for details). For estimates of decoupling (see below), the boundary layer top is the level of the inversion base, and the free troposphere begins at the inversion top. Two variables that have been argued to modulate Sc breakup are the jumps across the inversion in temperature T and specific humidity q (e.g., [Randall 1980](#); [Lock 2009](#); [Kawai et al. 2017](#)). The jump in specific humidity Δq is taken to be the difference between the value at the inversion base and top. For the temperature jump ΔT , to account for the fact that inversions are sometimes gradual, we consider the difference in temperature if a parcel at the inversion top is brought down to the inversion base along a moist adiabat.

Decoupling is estimated following [Wood and Bretherton \(2004\)](#), and LTS/EIS following [Wood and Bretherton \(2006\)](#). Both LTS and EIS are very similar metrics. We concentrate on using EIS as it is slightly closer to the physical state variable of interest (the strength of the MBL top inversion), though LTS is used to allow comparison to previous literature. For decoupling, though there are a number of comparable metrics with similar qualitative behavior, here we use the α_{qt} (total water alpha parameter), which varies from 0 (perfectly coupled) to 1 (upper MBL q_t is equal to FT q_t) and measures the

degree of stratification of the q profile (see [Wood and Bretherton \(2004\)](#) for a more in-depth definition). In a few cases where the liquid water content was unavailable due to instrumentation issues, only water vapor is used. This has a very minor effect on the decoupling estimates as the vast majority of MBL atmospheric water is vapor.

Subsidence is calculated from ERA5 as follows: for each trajectory in a case, the reanalysis inversion layer is estimated. A time series of subsidence between outbound and return sampling is extracted as the mean value height-averaged over this layer, from which a 2-day mean is then calculated. Since this is entirely based on reanalysis, potentially poorly constrained, and a key variable for our discussion, we include an estimate of the uncertainty in this mean subsidence value by making use of the ERA5 10-member ensemble, as well as the fact that we potentially have multiple trajectories per case. Therefore, the distribution of 2-day mean subsidence values over the ensembles and trajectories is calculated, and the standard deviation of this is taken as the uncertainty in the mean. The mean entrainment rate for each case is calculated by adding the subsidence estimate that to the mean rate of change of the boundary layer depth calculated using initial and final z_i estimates from the aircraft profiles (and therefore the subsidence uncertainty can be taken as a reasonable lower bound on the entrainment rate uncertainty). SST, surface fluxes, and 700 hPa free-tropospheric RH are taken from ERA5 reanalysis and averaged along the entire set of flight sequences making up a case.

Cloud-base rain rates are estimated by combining HCR and HSRL data collected during surface legs based on the technique proposed by [O'Connor et al. \(2005\)](#), as detailed in [Schwartz et al. \(2019\)](#). The precipitation rate for each leg was taken as the mean of column-maximum precipitation from all below-cloud legs, the height of which typically corresponds well with cloud base, typically between 400 and 800 m MSL. Columns with no precipitation were included as zero values (the precipitation rate is not rain conditional). The flight height of 500 ft (150 m MSL) meant that the cloud base was usually above the radar dead zone (<200 m from the plane).

3. Results

a. Characteristic Lagrangian evolution

In this section results are compared across all 18 cases considered together to look for robust features of the Sc–Cu transition. We derive mean changes averaged over all cases to define what a “characteristic” 2-day evolution

TABLE 1. Means of key variables for two case studies and over all 18 cases. Shown are outbound/return values and differences, as well as mean values along trajectories (bottom 3 rows). Values from cases are bolded if they fall outside the middle 80% of all values.

	Units	L06 (RF06BC/RF07A)			L10 (RF10A/RF11DE)			Mean over 18 cases		
		Out	Back	Δ	Out	Back	Δ	Out	Back	Δ
CF	%	95	39	−56	100	78	−22	84	53	−31
EIS	K	5.7	3	−2.7	8.5	3.6	−4.9	6.5	4.3	−2.2
LTS	K	19.9	16	−3.9	21.7	16.4	−5.3	19.3	17.1	−2.2
N_d	cm^{-3}	13	20	+7	227	43	−184	84	70	−14
z_i	m	750	2370	+1620	1240	1700	+460	1240	1800	+560
α_q	0–1	0.13	0.34	0.21	0.33	0.47	+0.14	0.23	0.35	+0.12
N_d	cm^{-3}	50	26	−24	190	207	+17	121	107	−14
prec	mm day^{-1}	2.3	0.5	−1.8	2.7	1.2	−1.5	2.2	0.7	−1.5
O_3	ppbv	15	15	0	27	33	+5	21	22	+1
CO	ppbv	70	70	0	118	117	−1	86	83	−3
SST	K	294	299	+4.6	294	297	+3.1	294	298	+3.5
FT RH	%	43	18	−25	13	37	+24	31	21	−10
ΔT	K	4.0	1.1	−2.9	5.3	2.6	−2.7	4.6	3.7	−0.9
Δq	g m^{-3}	−1.9	−5.2	−3.3	−2.4	−3.2	−0.8	−4.4	−5.3	−0.9
		Mean along trajectories			Mean along trajectories			Mean along trajectories		
Subsidence	mm s^{-1}	0.23 $\pm$ 0.6			3.5 $\pm$ 0.6			2.1 $\pm$ 0.8		
Entrainment	mm s^{-1}	9.5			5.9			5.2		
LHF	W m^{-2}	125			105			109		

of the subtropical cloud-topped MBL might look like. These values are shown in the final column of Table 1 as well as highlighted in Figs. 3 and 4.

1) LAGRANGIAN COHERENCE OF MBL PROPERTIES

As CSET resampled air masses after two days, we can examine the correlations of various variables sampled on the outbound and return flights to analyze the persistence or Lagrangian coherence of anomalies. Assuming a perfect resampling of an air mass, the correlation of outbound with return values should indicate how well-preserved anomalies are over time. We first look at the Lagrangian coherence of long-lived chemical tracers (ozone and carbon monoxide). Figures 3a and 3b show a strong correlation between outbound and return below-cloud measurements of both tracers for all 18 cases, which lends confidence to the Lagrangian measurement strategy. Deviations from a perfect correlation are expected due to entrainment into the MBL of air with different properties, chemical processing within the MBL, and air–sea fluxes. In particular, ozone in the lower troposphere has a relatively short lifetime of only a few days (Seinfeld and Pandis 2006), and so while it does not act as a perfect tracer, anomalies should persist long enough to correlate after 2 days.

A similar analysis on MBL properties of depth, decoupling, cloud fraction, cloud droplet number concentration, precipitation rate, and aerosol concentration shows relatively low Lagrangian coherence (Figs. 3c–h). This analysis is consistent with the finding of Eastman et al. (2016) that

these properties have e -folding time scales of 12–24 h, much less than the two day resampling time. The aerosol number concentration is an apparent exception, but the correlation coefficient drops to 0.52 if one very high-aerosol anomaly is removed. The mean precipitation estimates for the cases were highly sensitive to the how precipitation was defined. A more thorough analysis of the precipitation observations from CSET, including comparison of in situ and remotely sensed values, an evaluation of changes in precipitation frequency as well as mean rain rate, and a discussion of the role of precipitation in the transition, can be found in Sarkar et al. (2019, manuscript submitted to *Mon. Wea. Rev.*). In particular, they find that in situ observations of precipitation rate are higher than those estimates from the radar-lidar retrievals.

We conclude that while it is simpler to think about MBL evolution following air columns, as it matches the conceptual model of an air mass subjected to changing forcings, that evolution is in practice no more coherent than the spatial gradients of MBL and cloud properties between CA and HI.

2) TRANSITION FEATURES

We use two metrics to describe composite changes across our set of Lagrangian cases. The first is the mean difference in a variable between the outbound and return sampling over all cases (e.g., cloud fraction on average decreased 31%, from 84% to 53%), and the second is the number of cases experiencing an increase/decrease in a variable (e.g., 16 out of 18 cases experienced a decrease in cloud fraction. Note that the total number of cases can

FIG. 3. Lagrangian coherence of (a) ozone, (b) carbon monoxide, (c) inversion height, (d) decoupling index, (e) GOES cloud fraction, (f) cloud droplet number concentration N_d , (g) cloud-base precipitation rate, and (h) accumulation mode aerosol number concentration N_a . The x axis is the outbound flight value, and the y axis is the return flight value. Also plotted is the 1–1 line (dashed gray; this is meant as a quick guide to whether values increase or decrease between samplings); r values for a linear best-fit are in the top left. Red numbers mark the two Lagrangian case studies discussed in section 3b (L6 and L10), and black stars mark mean value across all cases.

FIG. 4. MBL evolution over two days of (a) cloud fraction vs EIS, (b) decoupling vs depth, (c) cloud fraction vs cloud droplet number concentration, (d) accumulation mode aerosol number concentration vs cloud droplet number concentration, (e) precipitation vs droplet concentration, and (f) cross-inversion moisture jump vs cross-inversion temperature jump. Blue dots show values from outbound flights, and red show values from return flights. Thick black line with larger dots shows mean behavior. Red and blue numbers mark the two Lagrangian case studies discussed in section 3b (L6 and L10).

FIG. 5. (a) Expanded trajectory set with evolution over 3 days of (b)–(e) GOES-retrieved variables and (f) ERA5 EIS. Red lines show trajectories sampled during CSET, black lines show all summer 2015 trajectories. Plotted are mean values averaged over all trajectories (thick lines with dark shading) and bounding quartiles (thin lines). Trajectories are shifted in time such that the local solar times at start align. In (a), trajectories go from NE to SW, and a random 50-member subset of the expanded trajectory set is shown [full set used for (b)–(e)].

be affected by missing data). These data are also shown in Fig. 4, which shows the evolution of all 18 cases in terms of various variable pairs. Each panel shows each case as a line between outbound (blue) and return (red) values. This highlights both the mean behavior as well as case-by-case variability.

In general these two metrics consistently show that most air masses undergo a “classic” stratocumulus to trade cumulus transition, including MBL deepening (13/18 cases, mean inversion base [z_i] change is +560 m), increased decoupling (14/17, +0.12), decrease in cloud fraction (16/18, −31%), decrease in EIS (18/18, −2.2 K), and decrease in N_d (10/17, −14 cm^{−3}) (Figs. 4a–d). On average, the moisture inversion becomes more pronounced ($[\Delta q]$ decreases by 0.9 g m^{−3}, and 14/18 cases show a decrease), while the temperature inversion weakens ($[\Delta T]$ decreases by 0.9 K, 12/18 cases show a decrease) (Fig. 4f).

The downstream changes in precipitation rate and cloud droplet number concentration N_d (Fig. 4e) are much less consistent. For precipitation, 14/18 cases showed a decrease, but the mean change was only −1.5 mm day^{−1} with a large spread. The decrease in precipitation can be explained with reference to Bretherton et al. (2019), their Fig. 10c; the local minimum at −143°E along the CSET transect in precipitation corresponds with

the mean location of resampling. Translated into the longitude–prime coordinate used for compositing aircraft data in Bretherton et al. (2019), the initial samplings were on average at −134°E and the return samplings at −144°E. An in-depth analysis of the Lagrangian evolution of precipitation for a subset of CSET cases can be found in Sarkar et al. (2019, manuscript submitted to *Mon. Wea. Rev.*), which contains a thorough characterization of precipitation frequency and rate, and also compares in situ and radar-sensed precipitation estimates. For N_d , the mean change was a decrease of −14 cm^{−3}, consistent with the decrease in N_a , but this was dominated by a few initially polluted cases with large decreases in N_d , and in fact 9/14 cases showed a slight increase in N_d .

The ubiquity of the classic SCT evolution could partly be due to the sampling strategy in CSET. Flights were targeted toward low-level sampling of initially cloudy MBLs that were forecast to advect down the EIS gradient to the southwest. One way of assessing to what extent the CSET sampling strategy biased the sampled air mass population is by comparing satellite data along the CSET trajectories to those from the expanded trajectory set, as shown in Figs. 5a–e. The time offset of the CSET trajectories is due to the difference in initialization time, as trajectories from the larger set

were initialized at 0000 UTC while CSET trajectories were usually sampled around 1800 UTC (local morning) on both the outbound and return flights. CSET outbound flights preferentially sampled trajectories originating at the northern end of the sample area. Although we do not have data to assess the representativeness of the decoupling state or aerosol environment of the boundary layers sampled during CSET, we can compare the evolution of GOES-retrieved cloudiness, cloud-top height and N_d , and ERA5-inferred stability between the CSET and expanded populations of trajectories. Although all fields show high variability, there does not appear to be a strong bias in the CSET sampled trajectory set. In particular, due to the filtering of the expanded trajectory set by direction, the evolution of EIS along both sets of trajectories is very consistent, indicating that the CSET sampled trajectories were not much different from a random set within the observational period, given the constraints on direction and initial location, though we cannot say whether they are more broadly climatologically representative given the strong possibility of significant interannual variability. Compared to the trajectory set used in Sandu et al. (2010) and subsequent work, the trajectories used here are farther to the northwest, with more of an easterly wind. They therefore experienced a weaker gradient in stability, which is consistent with the overall slower evolution in cloudiness in our set.

Figure 5 also gives some indication of the magnitude of the diurnal cycle of subtropical MBL cloud-top height and fraction. These values are obtained by first detrending and then fitting a sinusoid to the mean trajectory in Fig. 5, and the stated values are twice the fitted amplitude. Both maximize in the early morning and decrease during the day, cloud-top height by 150 m and cloud fraction by 13%. Nighttime deepening and infilling counteract both daytime trends, though the general behavior is still a net decrease in cloud fraction and deepening of the cloud layer toward Hawaii. The diurnal cycle of both depth and cloud fraction in marine stratocumulus is consistent with previous literature (e.g., Rozendaal et al. 1995; Burleyson et al. 2013; Painemal et al. 2013), and a more in-depth discussion of the diurnal cycle as it relates to the transition in CSET can be found in Sarkar et al. (2019, manuscript submitted to *Mon. Wea. Rev.*). It is worth noting that the aircraft observations, which all occur within a few hours of local noon, do not sample this diurnal variability. This expanded capability highlights the complementarity of the satellite and aircraft sampling. Care must be taken when examining the diurnal cycles of liquid water path and N_d . Though some of the measured variability is undoubtedly physical, there is also a known solar zenith angle bias

(SZA) in these retrievals (Grosvenor and Wood 2014). This is mitigated by filtering for SZA above 70° in Fig. 5.

3) CONTROLS ON EVOLUTION

Past work has explored the extent to which changes in MBL clouds are consistent among variables and whether such changes are predictable. We can validate some of that work using data presented here. The cloud fraction–stability relationship (as measured by EIS or LTS) is well studied, primarily in an Eulerian framework. Klein and Hartmann (1993) used geographical, seasonal and interannual changes in stratus cloud cover and LTS to estimate a $6\% \text{ K}^{-1}$ sensitivity of marine low stratus to LTS; Wood and Bretherton (2004) showed a similar sensitivity for EIS. Here we calculate the EIS/CF relationship using the CSET Lagrangian cases. The slope of the “mean Lagrangian” line (thick line) in Fig. 4a, representing the average behavior or the cases in CF/EIS space, is $14\% \text{ K}^{-1}$, steeper than those past climatological sensitivity estimates. Alternatively, we can consider the range of CF/EIS slopes for all 18 cases, which have a median and standard deviation of 15% and $21\% \text{ K}^{-1}$. CSET deliberately targeted Sc farther offshore of persistent coastal clearing (low CF) due to offshore flow over cold water (high EIS). Inclusion of such regions would likely lower the mean sensitivity more into line with the climatological sensitivity. Another possible explanation for the slope discrepancy is uncertainty in the CSET slope estimate, since it has a large range across individual cases. However, our derived sensitivity compares well to an estimate from a similar satellite study by Sandu et al. (2010) that considered approximately 8500 trajectories in the NEP. Based on their Figs. 2 and 3, considering trajectory evolution from day 1 to 3 so that the starting cloud fraction is closest to our sample, the mean cloudiness decreases 55% and LTS decreases 3.7 K over two days, implying a sensitivity around $15\% \text{ K}^{-1}$. This is despite the weaker LTS gradients in our sample.

The relationship between decoupling and depth is also strong, as shown in Fig. 4b. The slope of the mean trajectory behavior (thick line) is approximately 0.21 km^{-1} (i.e., the decoupling index increases by 0.21 for every km of deepening). The estimate of this slope as provided in Wood and Bretherton (2004) is 0.28, within estimation error (we used a bootstrap method to generate 1000 samples of 18 cases and estimated the mean slope as $0.22 \pm 0.06 \text{ km}^{-1}$).

Although the aircraft resampling only allows for studying of 2-day lagged correlations, the satellite and reanalysis data along trajectories has much higher time resolution, so it can give more insight into the time scales of evolution. Figure 6 shows the lag correlation

FIG. 6. Lag correlations of EIS and GOES cloud fraction/cloud-top height. Peak correlation for CF/EIS was at roughly -30 h lag (i.e., upstream EIS with downstream CF). For CTH, instantaneous correlations were strongest.

of GOES CF with EIS using the expanded trajectory set. Each hour represents the correlation among all trajectories between EIS lagged that many hours behind/ahead of the GOES CF 2 days into the trajectory (e.g., the 0-h lag correlation is the instantaneous correlation of EIS and CF 2 days after trajectory initialization). Also shown are lag correlations replacing CTH for CF. CF correlates best with upstream EIS 30 h earlier, strongly suggestive of EIS control on CF and consistent with Klein et al. (1995). No such lag exists in the EIS–cloud-top height correlations, though whether this is due to a more immediate response of the MBL depth to large-scale meteorology is uncertain.

As both entrainment and precipitation are important processes in MBL evolution, and aerosol can affect both of these through modulation of the cloud droplet number concentration, there is potential for strong aerosol influence on MBL evolution. The CSET data shows a strong correlation ($r = 0.72$) between below-cloud N_d and in-cloud N_d (Fig. 4d). However, there is no evidence of a significant aerosol influence on precipitation or entrainment rates. This is most likely due to the small sample size and the resulting inability to control for additional confounding factors such as MBL depth and large-scale meteorology, and not a general conclusion about the existence of indirect aerosol effects. Although a positive correlation was found between GOES N_d and the GV CDP N_d measurements (see Bretherton et al. 2019), use of the expanded trajectory set and their GOES N_d retrievals to investigate cloud lifetime effects is problematic. One reason for this is a possible cloud fraction-dependent bias related to the validity of the plane-parallel, adiabatic, or lapse rate assumptions applied in the retrieval of N_d for cumulus clouds. Correlations between N_d and GOES cloud fraction disappear when considering aircraft N_d instead of GOES, and

so we cannot state conclusively where there is an aerosol effect on cloud evolution present in the CSET Lagrangian cases.

b. Lagrangian case studies

In this section, we take a closer look at two Lagrangian cases observed in CSET, both initially similarly cloudy. L06 is shallow, and relatively clean, while L10 is deeper, more decoupled, and more polluted. The “slowly evolving” L10 case remains deep, decoupled, and cloudy, with little development toward a cumulus-dominated boundary layer, while the “rapidly evolving” L06 case experiences a rapid transition from stratocumulus into cumulus cloud involving deepening and decoupling. We explore the differing evolution in these two cases and document them to encourage their future use for process modeling studies.

In L06, cloud fraction decreases from 95% to 39%, the MBL deepens from 750 to 2370 m, and the decoupling parameter α_q increases from 0.13 (well-coupled) to 0.34 (moderately decoupled). In contrast, in L10, cloud fraction decreases from 100% only to 78%, the MBL slightly deepens from 1240 to 1700 m, and α_q goes from 0.33 to 0.47 (moderately to heavily decoupled). These cases are marked as “6” and “10” in Figs. 3 and 4 from the previous section. Table 1 highlights key variables from these two cases, as well as the average values over all 18 cases. Values that fall in the top or bottom decile of all 18 cases are bolded to highlight ways in which these cases were unusual.

1) METEOROLOGICAL ENVIRONMENT

Much of the difference between these two cases may be attributable to their meteorological environment and MBL forcings. Though it would be ideal to have two cases with only one crucial difference (e.g., aerosol environment), that is not achievable in a nature, and so we explore the large-scale forcings on these two cases first.

L06 was initially stable and experienced a moderate decrease in EIS. L10, though starting with a higher EIS, experienced a strong decrease and ended up with a similar EIS to L06. This can be seen in Fig. 4a, the L06 line is much steeper than the L10 line. The free-troposphere relative humidity was initially higher in L06 but ended higher in L10, such that the two cases experienced entirely opposite trends in FT RH (see Figs. 7b,e). L10 is unusual within our sample of cases, with its initial and final FT RH being anomalously low and high, respectively.

An important difference between these two cases is the large-scale vertical motion (Fig. 8d). L10 had a strong mean subsidence rate at the inversion (from ERA5) of approximately $3.5 \pm 0.6 \text{ mm s}^{-1}$, while L06

FIG. 7. Soundings and relative humidity cross section from L06 and L10. (a),(d) RH and theta from aircraft initial soundings along with inferred inversion base height (dotted lines; also marked on center panels with arrows). (b),(e) Evolution along trajectories (L06: trajectory 2.3; L10: trajectory 6.0) of RH from ERA5. RH curtain (RH as function of height and time) is only from one trajectory, not averaged over all case trajectories, for clarity. Toward the start and end of the trajectory (left and right ends of each panel), the ERA data are blanked out with a gray box and the aircraft RH data are overplotted, showing generally good agreement with ERA5 with regards to the height of the strongest RH gradient. White line marks GOES cloud-top height (75th percentile from $2^\circ \times 2^\circ$ box along trajectory); line is dotted where $CF < 50\%$. (c),(f) As in (a),(d), but for return soundings.

very weak subsidence of only $0.2 \pm 0.6 \text{ mm s}^{-1}$, consistent with its rapid deepening and decoupling. This plausible conclusion has two major sources of uncertainty. First, we do not have an independent test of the realism of the large oscillations in the ERA5 vertical motion following the trajectories in both cases, let alone their time average. Second, there was a strong vertical gradient in ERA5 vertical motion during L06, which makes the vertical velocity at the inversion base sensitive to uncertainties in how the inversion base is estimated. In fact, subsidence at 700 hPa averaged along the trajectories was comparable for both cases (between 3 and 3.6 mm s^{-1}).

The entrainment rate for the two cases is estimated by subtracting the vertical motion at the inversion base from the inversion deepening rate. Using the inversion heights from aircraft profiles (Fig. 7), the mean entrainment rates for L06 and L10 were 9.5 and 5.9 mm s^{-1} (± 0.6 , compared to an average across all cases of 5.2 mm s^{-1}). The very high L06 entrainment rate, in addition to weak subsidence, contributed to its rapid inversion deepening. Figures 7a and 7c show that the

L06 inversion height change from the aircraft profiles is higher than what GOES cloud-top height or ERA5 RH might indicate (in particular considering the inversion height at the start of L06), and a more conservative estimate based on visual inspection of Fig. 7 would be closer to 7 mm s^{-1} . Regardless, L06 shows stronger entrainment, likely due to the weaker inversion, even when considering the uncertainty in subsidence estimates.

Figures 7b and 7e also compare the GOES cloud-top height (white line), the aircraft inversion (colored triangles) and the ERA5 humidity inversion (transition between purple and orange shading). Note that for clarity, the ERA RH curtain is only drawn from one trajectory per case and not averaged over all trajectories. This is for a more accurate portrayal of the ERA5 vertical structure. When the MBL is deep and cloud fraction is low (marked by dash in GOES cloud-top height), much of the cumulus cloud can be seen through the disappearing deeper stratus and the GOES cloud-top height drops well below the other inversion base estimates (consistent with Karlsson et al. 2010),

FIG. 8. Evolution of cases along trajectories with aircraft validation: (left) L06 and (right) L10. Each line represents data along a different trajectory for that case, and red dots mark aircraft data with interquartile range. (a) GOES low warm cloud fraction, (b) GOES 75th percentile N_d with aircraft CDP N_d , (c) ERA 700 mb specific humidity, (d) ERA subsidence averaged over inversion layer, (e) ERA 1000 mb wind speed with aircraft below-cloud leg (150 m) wind speed, (f) ERA SST with aircraft radiometric surface temperature from below-cloud leg, and (g) ERA surface latent heat flux. Gray shading indicates night time, and thicker lines (trajectories 2.3 and 6.0) are highlighted for comparison with other plots.

though a possible bias in GOES cloud-top height estimates in broken scenes due to surface contamination (Zuidema et al. 2009) has not been ruled out. This effect also has a strong diurnal signal from daytime stratocumulus burn-off, which complicates naive use of cloud-top height as a proxy for boundary layer depth.

The differences in surface latent heat fluxes (LHF) shown in Fig. 8 can be explained by considering the SSTs and near-surface wind speeds. The former was higher in

L06, consistent with its location slightly farther along the climatological SST gradient. This resulted in an initially higher LHF, though the decrease in wind speed toward the latter half of L06 resulted in the LHF dropping below the L10 values. Averaged over the two days between sampling, the L06 LHF was ultimately above average, and conversely L10 LHF was below average. As latent heat fluxes are a key driver of MBL decoupling (Bretherton and Wyant 1997), the increase in L06 decoupling and

relative lack of further decoupling of L10 is consistent with these LHF anomalies.

2) AEROSOL ENVIRONMENT

The aerosol environment also differs between the two cases. Relative to the ensemble of cases, L06 was unusually clean while L10 was unusually polluted by every measure, including initial N_a and N_d (as shown in Fig. 4d), CO, and O₃ (as shown in Figs. 3a,b).

There are a number of potential consequences of the more polluted environment of L10 on the cloud evolution. According to LES, an initially higher N_d should increase entrainment while decreasing cloud-base precipitation, with an uncertain net effect on cloud liquid water path (Stevens et al. 1998; Ackerman et al. 2004). Though initially under a free-troposphere dry relative to the case average, the elevated FT RH toward the second half of L10 (Figs. 7e and 8c) preferentially pushes the MBL toward a regime where Sc can thicken and the precipitation effect should begin to dominate (Ackerman et al. 2004; Yamaguchi et al. 2017). However, observed precipitation rates were comparable in L10 to L06, and the entrainment rate was higher in L06 rather than L10. This suggests that confounding meteorological differences between L10 and L06 are overwhelming potential aerosol effects. Sarkar et al. (2019, manuscript submitted to *Mon. Wea. Rev.*) investigated differences in precipitation in the flight data in more depth and found that the precipitation reaching the surface was greater in the L06 case, consistent with the cleaner boundary layer.

The N_a/N_d coevolution in case L10 (Fig. 4d) shows a very strong decrease in N_d with no significant change in N_a . This can be explained by considering the flight legs from which these measurements were taken and degree of MBL decoupling. Aerosol measurements are from the 150 m subcloud leg (to avoid contamination by liquid water), while N_d measurements are from the in-cloud leg, which in the return flight from L10 were taken in the upper decoupled cloud layer (at 1550 m). The decrease in N_d is likely explained by the cleaning of this decoupled layer through droplet scavenging by precipitation, together with the decoupling from the lower mixed layer aerosol reservoir (Wood et al. 2011). This was confirmed by comparing the L10 downwind N_a observations from the cloud layer (filtered to remove samples containing liquid water) and the subcloud N_a observations: cloud-layer N_a was 114 cm^{-3} and subcloud N_a was 207 cm^{-3} , still elevated above the N_d of 43 cm^{-3} but not as extreme a difference. We do not use the cloud-layer N_a throughout since the required cloud filtering results in too few data points in many cases, and so the decoupling state must be kept in mind. From the observations we cannot support a cloud lifetime effect of aerosol to explain the slow breakup in L10.

In short, the slowly evolving, polluted L10 case, despite a strong reduction in stability, did not deepen or experience much decrease in CF. This is possibly a result of lower surface fluxes, stronger subsidence, and suppression of an aerosol effect from high free-tropospheric relative humidity. That said, as we have shown, there is a lag in the response of cloud fraction to the effect of decreasing stability, and Fig. 8a shows some evidence that breakup became more prominent in the 18 h after resampling. In contrast, L06 had a weaker inversion and higher average surface fluxes, resulting in greater entrainment, which, together with the significant subsidence, allowed for rapid deepening and decoupling. This deepening, combined with a cleaner aerosol environment, likely resulted in vigorous precipitation and further cleaning of the boundary layer and depletion of the Sc layer.

4. Summary and discussion

In this work we have presented the CSET campaign data organized into 18 Lagrangian cases. These cases present an excellent opportunity for future modeling studies and greatly expand the number of possible aircraft-based modeling cases. We have shown the construction and context of these cases, as well as the limitations in their interpretation, and presented an in-depth look at two case studies.

Our analysis of the CSET Lagrangian sampling provides some useful insight into the nature of MBL cloud evolution in the Sc–Cu transition. Many of the trajectories were initially sampled in Sc, but downwind resampling after two days showed a Lagrangian transition to cumulus clusters associated with higher SST, weaker inversion stability, and deeper, more decoupled MBL structure, in accord with existing theory (Bretherton and Wyant 1997). A diversity of aerosol regimes was sampled, but the effect of aerosol variations on MBL evolution was not easy to separate from other meteorological factors.

Results from analysis of the trace gases CO and O₃ validate the CSET HYSPLIT Lagrangian track-and-resample flight strategy, even in the face of complications such as vertical shear within the MBL and inexactness of deriving constant-level trajectories from operational analyses. Despite the good resampling, most key cloud and MBL fields had decorrelated at the 2-day resampling, consistent with the satellite-based Lagrangian analysis of Eastman et al. (2016). Variability in fields that do not follow the boundary layer mean flow is likely responsible. In particular, rapid changes in vertical velocity along boundary layer trajectories will affect cloud thickness and thereby precipitation and aerosol removal, causing

quick decorrelation in those fields. The CSET sampling strategy may further lower the measured Lagrangian correlation, as each Lagrangian case spans a large range of mesoscale variability (with only intermittent sampling from either profiles or certain level legs). Fields with higher spatial heterogeneity such as precipitation or N_d will be more strongly affected. This limitation should be considered for future Lagrangian aircraft-based studies.

Geostationary satellite data along the Lagrangian trajectories were analyzed to complement the aircraft sampling and more completely document MBL evolution. Aircraft data provide definitive and complete cloud, aerosol, precipitation and thermodynamic profiling, while satellite data provide good space–time coverage for selected variables subject to potential retrieval caveats. The context of satellites and modern global meteorological reanalyses immensely expands the science that can be done with a single research aircraft used in a survey mode as in CSET, and the use of geostationary satellites for Lagrangian studies in particular is effective due to the high temporal sampling they offer. As with many prior studies we found a significant Lagrangian correlation between the evolution of EIS and cloud fraction, maximizing when cloud fraction is compared with the EIS seen along the same trajectory 30 h prior, consistent with previous literature (Klein et al. 1995). Using the expanded trajectory analysis, we showed that the boundary layers sampled by CSET were representative of the region (though interannual variability limits the extent to which this season can be considered climatologically representative) and quantified the strength of the diurnal cycle of CF and CTH.

A motivation of the CSET campaign was to provide datasets suitable for process modeling cases, for which a Lagrangian framework is advantageous because it reduces uncertainties associated with horizontal advective tendencies. The Lagrangian evolution of the MBL was quite variable from day to day. Influences on MBL evolution are strongly state dependent (e.g., the effect of aerosol on stratocumulus entrainment depends on whether the cloud layer is thin or thick). If a composite case is created for model comparison (e.g., de Roode et al. 2016), much of this complexity is lost and indeed it can be challenging to make quantitative comparisons with observations. By providing a variety of potential Lagrangian modeling cases with detailed aircraft observations, our CSET data should provide modelers with more options for exploring MBL evolution. In this paper, these cases have been placed into context, and two cases have been described in detail, which span some of the diversity of SCT evolution observed during CSET. The cases differ in their initial depth, decoupling,

microphysics, and large-scale forcings. We plan to present large-eddy simulations of these cases in a forthcoming paper using CSET aircraft data for initialization and comparison after 2 days, with reanalysis extracted along trajectories providing the large-scale forcings, and we hope future researchers will take an interest in which of these factors is most important to the cloud evolution. Most of our other 18 Lagrangian cases could also be fruitful for comparison with process models, as they sample a diversity of meteorological and aerosol conditions. An important caveat is the inevitable oversimplification in considering a flight sequence spanning hundreds of kilometers as representing a single air mass, since it will invariably sample across mesoscale variability and spatial gradients. For instance, the representativeness of a particular aircraft profile for initializing or comparing with a Lagrangian modeling case must be considered.

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APPENDIX

Estimation of Inversion Properties from Profiles

Conceptually, the inversion capping the marine boundary layer is easy to understand as a layer at the top of the marine boundary layer marked by an anomalous temperature increase and moisture decrease with increasing height. In practice, there is significant variability in the presentation of the inversion layer. The presence of moist layers aloft, differences in background free-tropospheric humidity and temperature, and residual decoupled layers with their own inversion complicates

FIG. A1. Example profiles from CSET showing estimation of inversion layer (green lines across all plots, marking base, center, and top of the inversion) and ΔT , Δq (red lines on plots of temperature and moisture). (a),(e) Temperature inversion; note that the inversion top temperature is adjusted to the altitude of the inversion base along a moist adiabat. (b),(f) Moisture profile and (c),(g) relative humidity. Note the presence of an elevated moist layer in (g) well above the strong temperature inversion in (e). (d),(h) (smoothed) Lapse rate and moist adiabatic lapse rate.

finding a single working definition. Additionally, the vertical extent of the inversion varies significantly from the usually strong and vertically compact inversions in the shallow stratiform MBL regimes closer to CA to deep and gradual inversions in the trade regions downwind. When considering multiple data sources, such as the reanalysis and aircraft profiles considered in this work, differences in data resolution add further complexity. Other measures of the depth of the boundary layer, such as those based on cloud-top heights, have their own complications beyond the scope of this appendix. In this appendix we will define our measure of the inversion layer (with a base, center, and top) and how we estimate ΔT and Δq across the inversion. We will also discuss how aircraft data inversion estimates compare with reanalysis, and how they compare with other methods of the inversion height.

Preprocessing of aircraft profile and dropsonde data are required; we first smooth with a 30-s sliding Hanning window (corresponding to a vertical extent of roughly 220 m for aircraft profiles and 180 m for dropsondes). Our approach is as follows (see Fig. A1 for a visual aid): 1) calculate the “center” of the inversion by identifying the region of strongest anomalous stability (strongest difference between actual lapse rate and moist lapse rate, $\Gamma - \Gamma_m$), 2) calculate the top of the inversion as the lowest level above the center where $\Gamma > \Gamma_m$, then 3) calculate the lowest extent of the inversion as the highest level below the center where $\Gamma > \Gamma_m$. This isolates the region of anomalous stability well, with the center of the inversion separating the upper and lower inversion regions. We define Δq as the difference in specific humidity between the top and bottom of the inversion. For ΔT , we integrate $(\Gamma - \Gamma_m)dz$ from the

inversion base to the inversion top. This is equivalent to bringing a parcel from the inversion top down to the level of the temperature minimum along a moist adiabat. A slight modification to the method above was needed to account for the occasional presence of a layer of slightly increased stability above sharp Sc inversions caused by strong radiative cooling. In step 2, the inversion top was calculated instead as where $\Gamma - \Gamma_m > \max(\Gamma - \Gamma_m)/4$ (i.e., where the lapse rate gets 75% of the way back from greatest deviation back to moist adiabatic). The same adjustment was made in step 3 for the inversion base to avoid including the entire cloud layer in more decoupled boundary layers. The exact cutoff is somewhat arbitrary but was found to work well for the boundary layers analyzed and falls between two reasonable definitions of the inversion layer, the first being where the lapse rate is subadiabatic and the second being where the lapse rate is positive.

Figure A1 shows a couple aircraft profiles from CSET, marked with the relevant inversion layers (base, middle and top). It also shows the temperature and moisture values used in calculating ΔT and Δq . In Fig. A1b, note the discrepancy between the temperature inversion and the moisture profile; the presence of moist layers aloft confounds a moisture-based estimate of the inversion. Note also in both profiles the slightly subadiabatic layer extending 500 m above the diagnosed inversion top.

This measure was used primarily because it gives reasonably good agreement between aircraft and reanalysis estimates of the inversion center height ($r = 0.74$, mean bias of 29 m with higher ERA5 inversions, though the thickness of the inversion layer is consistently overestimated in reanalysis), and because it agrees well with the Heffter method used in prior work (Heffter 1980; see Bretherton et al. 2019 and McGibbon and Bretherton 2017) on aircraft and dropsonde profiles ($r = 0.85$, mean bias of 89 m in with lower Heffter inversion heights). For ERA5 comparisons, the grid point nearest the middle of the aircraft/dropsonde profile is selected and no smoothing is carried out.

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