Observed Boundary Layer Controls on Shallow Cumulus at the ARM Southern Great Plains Site

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ABSTRACT

The boundary layer controls on shallow cumulus (ShCu) convection are examined using a suite of remote and in situ sensors at ARM Southern Great Plains (SGP). A key instrument in the study is a Doppler lidar that measures vertical velocity in the CBL and along cloud base. Using a sample of 138 ShCu days, the composite structure of the ShCu CBL is examined, revealing increased vertical velocity (VV) variance during periods of medium cloud cover and higher VV skewness on ShCu days than on clear-sky days. The subcloud circulations of 1791 individual cumuli are also examined. From these data, we show that cloud-base updrafts, normalized by convective velocity, vary as a function of updraft width normalized by CBL depth. It is also found that 63% of clouds have positive cloud-base mass flux and are linked to coherent updrafts extending over the depth of the CBL. In contrast, negative mass flux clouds lack coherent subcloud updrafts. Both sets of clouds possess narrow downdrafts extending from the cloud edges into the subcloud layer. These downdrafts are also present adjacent to cloud-free updrafts, suggesting they are mechanical in origin. The cloud-base updraft data are subsequently combined with observations of convective inhibition to form dimensionless “cloud inhibition” (CI) parameters. Updraft fraction and liquid water path are shown to vary inversely with CI, a finding consistent with CIN-based closures used in convective parameterizations. However, we also demonstrate a limited link between CBL vertical velocity variance and cloud-base updrafts, suggesting that additional factors, including updraft width, are necessary predictors for cloud-base updrafts.

1. Introduction

Shallow cumulus (ShCu) clouds are an important component of the climate system, impacting the surface energy budget and thus planetary boundary layer development in both continental and oceanic settings. Despite their importance, ShCu remain incompletely represented in global climate models (GCMs) because of their subgrid-scale size (~ 1 km) and uncertainties in their relationship to resolved-scale variables (i.e., parameterization). As a

result, GCMs suffer from errors in the location, timing, and extent of both shallow and deep convective clouds (Dai and Trenberth 2004). Errors from these convective processes subsequently affect other components of the simulated climate system including tropical oscillations (e.g., Madden-Julian oscillation; Hannah et al. 2015) and structure of the planetary boundary layer (Park and Bretherton 2009). As such, there is a clear motivation for improving the physical and observational basis for parameterizing shallow convection in order to enhance the overall performance of GCMs.

In nature, ShCu form at the top edge of the convective boundary layer (CBL) and are typically associated with fair weather conditions but are also present on days with deeper convection. Individual cumuli develop from boundary layer updrafts that penetrate to the lifting condensation level (LCL), though variations in updraft water vapor can cause small local variations in cloud-base height (Crum et al. 1987). Climatological studies of

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continental ShCu indicate mean horizontal scales of about 1 km (Zhang and Klein 2013, hereafter ZK13; Lamer and Kollias 2015, hereafter LK15) and depths that vary from less than 100 m (e.g., cumulus humilis) to more than 1 km (e.g., cumulus mediocris).

To this end, ShCu can be classified as either *active* or *forced* based on their vertical extent and internal dynamics (Stull 1985). Forced clouds, for example, are typically less than 300 m deep and serve as tracers of boundary layer updrafts as they penetrate to the LCL (Stull 1985; ZK13). Active clouds, on the other hand, release moist instability as the initial updraft reaches the level of free convection (LFC), thereby allowing the cloud to generate its own buoyant ascent. Active clouds are typically more than 300 m deep and almost always exist as part of an ensemble of clouds that includes both active and passive members (LK15). An additional cloud category, *passive* or *decaying* clouds, describes clouds that are in a dissipative (e.g., evaporative) phase. The base of decaying clouds may not correspond with the ambient LCL (Stull 1985).

Since ShCu of all depths have their roots in the CBL, it is important to understand the interactions and feedbacks between cloud development and CBL processes. Comparing clear-sky and ShCu-topped CBLs, Nicholls and LeMone (1980) found minimal differences between virtual heat flux profiles due to cumuli. Similarly, Hogan et al. (2009) found insignificant impact of cumuli on the vertical velocity variance and skewness profile. In contrast, Chandra et al. (2010), using profiling radar and a multiyear dataset, found that ShCu-topped CBLs had lower variance (and therefore weaker turbulence) than clear-sky profiles, though among cloudy periods, those with higher cloud fractions had relatively higher variance. Ansmann et al. (2010), using Doppler lidar, found that the frequency of up- and downdrafts near cloud base was enhanced by a factor of ~ 1.5 during ShCu conditions, though their sample included only 3 days.

The factors controlling ShCu vertical development and spatial coverage have also been the focus of recent observational investigations (ZK13; LK15). ZK13 found that active ShCu preferentially form on days with higher boundary layer humidity and lower cloud-layer stability. Combined, these factors favor earlier cloud onset, lower cloud bases, and greater cloud water. LK15 found similar environmental sensitivities and further showed that active cumuli are 2–3 times wider than passive clouds, are linked to wider and stronger updrafts in the subcloud layer, and are associated with more positively skewed distributions of vertical velocity.

Interactions of the cloud layer with the CBL have also been addressed in a modeling and parameterization framework. Mixed-layer models show, for example, that

cloud-base mass flux and compensating clear-sky downdrafts retard CBL growth after cloud initiation (van Stratum et al. 2014). In addition, the downdrafts that compensate the updrafts within ShCu provide a tight coupling between cloud-base height and the CBL-top height by increasing the convective inhibition (CIN), which is a measure of the energy barrier between the CBL top and the LCL (or LFC, depending on how it is defined; Bretherton et al. 2004). This feedback process is the premise for closure assumptions in “CIN based” shallow convection parameterizations currently employed in some numerical models.

In a CIN-based closure, the ShCu updraft fraction and cloud-base mass flux are governed by the ratio of CIN to the CBL updraft kinetic energy (Bretherton et al. 2004; Park and Bretherton 2009). This closure assumption has been shown to apply not only to ShCu but also to deep precipitating convection (Kuang and Bretherton 2006; Fletcher and Bretherton 2010; Hohenegger and Bretherton 2011). Recently, however, other numerical investigations have suggested that even after the updraft energy exceeds the energy barrier strength, additional factors, including updraft width, affect cloud development (Rochetin et al. 2014a,b). These factors have not, however, been examined in observational studies.

The goal of this paper is to refine our understanding of CBL and ShCu interactions using a multiyear observational dataset. Specifically, we employ a “large-eddy simulation (LES)-like” suite of active remote sensing observations to examine the relationships among surface forcing, boundary layer kinematics, cloud-layer stratification, and cloud extent. A key instrument in the study is a Doppler lidar, which has been shown in recent studies to provide rich information about both CBL and cloud-base properties (Hogan et al. 2009; Ansmann et al. 2010; LK15; Berg et al. 2017). As part of our analysis, we also examine the applicability of the CIN-based closure assumptions for updraft fraction and cloud-base mass flux based on the observed distribution of CBL and cloud properties.

2. Data and methods

a. Location, instruments, data streams, and postprocessing

Data from collocated in situ and remote sensors at the Department of Energy (DOE) Atmospheric Radiation Measurement (ARM) Southern Great Plains (SGP) site are used. ARM SGP, located in north-central Oklahoma, is an ideal site for this study because of a high climatological occurrence of days with locally forced shallow continental convection. Accordingly, SGP has been the focus in a number of recent ShCu observational studies (Berg and Kassianov 2008; Chandra et al. 2010;

Zhang and Klein 2010; ZK13; LK15) and is currently the focus of intensive modeling for cloud-scale processes.

The SGP instruments used in this study along with their application and data processing are as follows:

- 1) Doppler lidar (DL): The DL is an active ground-based laser remote sensor that provides range- and time-resolved profiles of radial velocity and attenuated backscatter coefficient (backscatter) over the lower troposphere. The lidar employs a 1.5- μm (near IR) laser, 30-m range-gate resolution, and 1.3-s temporal resolution. The DL is most sensitive to micrometer-sized aerosol in “clear air” and rapidly attenuates in liquid water. It is thus ideally suited to study boundary layer vertical velocity as well as cloud-base properties. Two separate DL data streams are employed in this study. First, postprocessed DL data obtained from the ARM data archive (DLPROFWSTATS4NEWS in the ARM archive; <https://doi.org/10.5439/1178583>) are used to quantify the bulk CBL properties including higher-order moments of vertical velocity (VV), the temporal evolution the CBL depth, and the cloud fraction. This “value-added product” (VAP) computes the higher-order moments of vertical velocity for 15-min windows. As part of the processing, the contribution of instrument noise to the total variance has been removed using an auto covariance technique (Lenschow et al. 2000). There is a known issue with the DL profiles that produces an oscillatory artifact in some of the variance profiles prior to 2015 (D. Turner and R. Newsom 2016, personal communication). To examine the impact of this, we compared variance statistics prior to 2015 with an independent dataset of five DL now installed at and around ARM SGP and found that this artifact does not affect the conclusions reached in this paper, including the primary attributes of the variance profiles presented later. Additional DL-derived quantities in the VAP are the cloud-base height and cloud fractions for each 15-min period. The mixing height (which is analogous to the CBL depth during the daytime for our cases) is subsequently determined from the VV variance using a thresholding algorithm following Tucker et al. (2009) wherein mixing height is assigned to the first range gate where the variance drops below $0.08 \text{ m}^2 \text{ s}^{-2}$. If the threshold is never exceeded in the column, the mixing height is set to be the first lidar range gate. The mixing height, and our overall results, are relatively insensitive to changes in this value since the vertical gradient in variance is strong near the CBL top. Berg et al. (2017) recently applied a similar algorithm, but with an iterative thresholding approach, to study CBL properties across a yearlong dataset, also at SGP.

In addition to postprocessed DL data, the “raw” high-temporal-resolution DL data (DLPROF in the ARM archive) are used to identify individual cumuli and to quantify their subcloud circulations (e.g., cloud-base updrafts). These raw data are processed by the authors to remove data points with low signal-to-noise ratio ($\text{SNR} < 0.005$) and range filtered to eliminate large jumps ($> 8 \text{ m s}^{-1}$) in adjacent velocity observations. To be specific, each observation is compared to the eight adjacent observations in a time–height matrix of vertical velocity and is removed if the range threshold is met. The cloud identification algorithm leverages the bimodal distribution of attenuated backscatter coefficient wherein cloud backscatter belongs to a different population than aerosol backscatter (supplemental Fig. S1). Clouds are thus defined as temporally continuous vertical scans with backscatter above a threshold value ($-4.6 \text{ m}^{-1} \text{ sr}^{-1}$) separating the two populations. The algorithm further requires cloud-base height to be within 300 m of the CBL top and cloud duration greater than 30 s and less than 20 min. Short gaps in cloud returns ($< 20 \text{ s}$) are permitted before clouds are separated into individual occurrences. An example of the cloud identification process from the raw lidar data is shown in Figs. 1a and 1b. We further note that our cloud-base identification agrees well with the cloud-base height determined using Haar wavelet analysis of the backscatter profile, which is available as part of the lidar VAP. Our separate analysis of cloud occurrence was required because the lidar VAP does not include the start and end times of individual clouds.

Once identified, each cloud is assigned a representative cloud-base height, cloud duration, and a cloud chord length. The representative cloud-base height is defined as the 25th-percentile cloud-base height, determined from the 1-s data, for each cloud. The chord length is determined by averaging the two nearest Doppler lidar horizontal wind profiles [available from postprocessed lidar data from a velocity azimuth display (VAD) retrieval; `dlprofwind4news`; <http://dx.doi.org/10.5439/1190027>] to estimate the advective speed at cloud base. Next, we extract the lidar vertical velocity time series from 30 m below the representative cloud base in order to quantify cloud-base vertical velocity and mass flux (example in Fig. 1d). The total mass flux into the cloud (M_{cb}) is then computed as

$$M_{\text{cb}} = \rho \int_0^{\text{cl}} w_{\text{cb}} dl,$$

where w_{ick} is the vertical velocity 30 m below cloud base, dl is an increment of length, cl is the cloud chord length, and density is assumed to be 1 kg m^{-3} .

FIG. 1. Examples of the cloud identification and subcloud circulation analysis on 20 Jul 2011. (a) Attenuated backscatter from the Doppler lidar showing the structure of the CBL and the presence of a single ShCu cloud (apparent at yellow shaded values near 1120 CST). (b) The DL vertical velocity (red and blue shading) and cloud-base detections (black asterisks) for the same scene. (c) “Cloud scene” extracted from the lidar data in (b) wherein height is normalized by that of cloud base and time by cloud duration. (d) Cloud-base vertical velocity determined 30 m below the cloud base. The red line segments indicate updrafts ($w > 0.1 \text{ m s}^{-1}$). Note that the cloud corresponds to normalized times between -0.5 and 0.5 .

Mass flux data are only computed for clouds where more than 75% of the cloud base has valid observations. Invalid observations are those data points for which the signal-to-noise ratio is too small or the vertical velocity exceeds range thresholds. As part of the analysis, we also define the updraft fraction to be the fraction of the total cloud duration (which spans -0.5 to 0.5 on the normalized time scale) for which the vertical velocity exceeds 0.1 m s^{-1} . The mean (W_u) and 95th-percentile (W_{95}) updraft speeds along the cloud base are also recorded. Downdraft fraction and speed are defined analogously.

Finally, a subcloud “scene” of vertical velocity throughout the CBL is extracted for each cloud event. These data are normalized by cloud-base height and the cloud duration to facilitate composite analysis. Figures 1c and 1d show an example of the height- and time-normalized subcloud scene and cloud-base updraft data corresponding to the cloud identified in Figs. 1a and 1b. Note that the 25th-percentile cloud-base height (black stars are individual detections) corresponds to a normalized height of 1, and the cloud resides between normalized times -0.5 and 0.5 , such that it has unit duration centered on 0. This normalization facilitates composite analysis of clouds with different duration and height above ground.

- 2) Raman lidar (RL): The RL is an active, ground-based laser remote sensing instrument that measures vertical profiles of water vapor mixing ratio, cloud base, and

aerosol-related quantities. The RL transmits a 355-nm UV beam and receives the backscatter at multiple wavelengths including 408 and 387 nm, which are associated with the Raman scattered signals for water vapor and nitrogen, respectively (Turner and Goldsmith 1999). The postprocessed data at 10-min temporal resolution and 75-m vertical resolution (10RLPROFBE1NEWS in the ARM archive; <https://doi.org/10.5439/1027250>) are used to examine the relative humidity and mixing ratio in the CBL and cloud layer during ShCu days. These data are filtered using a cloud mask (derived from the RL depolarization ratio) to remove all cloudy points; thus, all reported humidity variables reflect clear-air values only.

- 3) Microwave radiometer (MWR): The MWR is a passive two-channel radiometer that retrieves column liquid water path (LWP) at 25–30-s intervals (Turner et al. 2007). We use the LWP as a proxy for cloud depth in some of our analyses, though we note that LWP varies for other reasons as well (MWRRET1LILJCLOU in the ARM archive; <https://doi.org/10.5439/1285691>).
- 4) Balloonborne sounding system (sonde): The sonde data provide in situ profiles of temperature, humidity, wind speed, and wind direction. The sondes are launched four times daily, but we use only data from the 1130 central standard time (CST; note that CST is UTC $- 6\text{h}$) launch to document the thermodynamic profile, including CIN as a measure of the energy barrier near the time of cloud initiation (sgpsondewnpn1.b1 in the ARM archive; <https://doi.org/10.5439/1021460>).

- 5) Atmospheric Emitted Radiance Interferometer (AERI): AERI is a ground-based passive sensor that measures the downwelling infrared radiance from Earth's atmosphere (Knuteson et al. 2004a,b). We use postprocessed data wherein the vertical profile of temperature and water vapor are retrieved at high temporal resolution (Feltz et al. 2003; AERIPROF3FELTZ in ARM archive; <https://doi.org/10.5439/1027279>). The temporal resolution of these data is ~ 8 min. Data are available at 50-m elevation increments, though the underlying resolution is closer to 1 km near the top of the CBL (Turner and Löhnert 2014). While these data provide useful information about the overall thermodynamic structure of the atmosphere and compare reasonably well against nonintegrated convective indices (not shown), the ability to resolve finescale stable layers, including the capping inversion at the top of the CBL, is more limited (Blumberg et al. 2017). As described below, these limitations do not appear to affect the conclusions of the analysis. For our analyses, AERI temperatures are subsequently converted to potential temperatures using the surface pressure and a hypsometric integration. The potential temperature data are then used to examine the temporal evolution of stratification within the cloud layer on ShCu days.
 - 6) Surface meteorology and fluxes: Surface meteorological observations (SMOS; in ARM archive) and quality-controlled eddy correlation flux measurements (QCECOR; in ARM archive; <https://doi.org/10.5439/1097546>) from SGP site E14 are used; E14 is very near the above mentioned remote sensors. The QCECOR data are used to characterize surface fluxes, and the SMOS data are used to compute the LCL as a function of time. The LCL is computed from the surface temperature and dewpoint temperature using the approximation $Z_{lcl} = 125(T_{dry} - T_{dew})$, which is within $\pm 2\%$ for typical meteorological conditions at SGP (Lawrence 2005).
- b. Event identification*
- Days with ShCu convection at ARM SGP are identified for the warm seasons (Apr–Sep) 2011–15 using the following criteria:
- 1) Qualitative assessment of ShCu development from visible satellite data (MODIS *Terra–Aqua*) to ensure local forcing and the absence of synoptic or mesoscale disturbances
 - 2) A well-developed CBL, determined from the DL vertical velocity variance observations and confirmed with surface heat flux data
 - 3) Cloud bases closely coupled (within 300 m) to the CBL top, as determined from the DL
 - 4) Minimal mid- and upper-level cloudiness, determined from DL attenuated backscatter profiles and satellite
 - 5) Completeness of available datasets
- Additional manual inspections of data are used to filter out days that are not representative of ShCu conditions (stratocumulus conditions in the morning, anomalous features in the vertical velocity data, etc.), and in total, 138 days are selected for the study.
- Figure 2 provides an example of one such representative ShCu day in terms of satellite and lidar observations. The visible satellite imagery [1410 CST from MODIS *Aqua* satellite] indicates a broad region of shallow cumuli spanning Oklahoma wherein cloud fraction and horizontal cloud size increase from northwest to southeast. The corresponding DL vertical velocity, derived CBL height, and cloud-base detections are shown in Fig. 2b and are illustrative of a typical ShCu day evolution: rapidly growing CBL in the morning, slower CBL growth in the afternoon, and cloud bases coupled to the CBL height and cloud fraction decaying in the evening.
- For reference, we also selected 108 clear-sky days to include in this study. The criteria for the clear-sky days are similar to those for the ShCu days with the obvious exception that cloud fraction due to CBL clouds remains nominally zero.

3. Results

a. Composite boundary layer

1) KINEMATICS AND CLOUDS

In this section, the composite structure and evolution of the CBL across the 138 ShCu days is examined (Fig. 3). All means and medians are computed for 15-min intervals. Figure 3a shows the composite time–height evolution of the CBL in terms of the median VV variance (gray shading), VV skewness (red contours), CBL height (solid black line), cloud-base heights and cloud fraction (color-filled circles), and the LCL (dashed green line). In these composites, the height normalization is with respect to the maximum CBL height for each day, which occurs at different times; thus, the normalized mean CBL height never reaches 1. For a more physical interpretation, we subsequently convert the normalized CBL height back to a physical height by multiplying by the mean of the maximum daily CBL heights (height above ground). Also shown are the mean surface sensible and latent heat fluxes (Fig. 3b), mean CBL and LCL heights (Fig. 3c), and mean cloud fraction (Fig. 3d). One standard error is indicated for each of these variables. Similar to the example in Fig. 2b, CBL

FIG. 2. Example of a ShCu day at ARM SGP [red dot in (a)]. (a) MODIS *Aqua* visible imagery showing the distribution of locally forced convective clouds at 1410 CST 4 Sep 2015. (b) Doppler lidar observations from 4 Sep 2015 at SGP showing the vertical velocity (color shading), derived convective boundary layer height (dashed black line), and cloud-base detections (green dots).

growth begins at ~ 0600 CST, advances rapidly through the morning, reaches a maximum height (in the mean) of 2082 m AGL at 1455 CST, and then decays in height during the late afternoon. The VV variance is greatest in the lower half of the CBL and reaches a maximum about 1 h prior to maximum CBL depth. The VV skewness is strongly positive throughout, which is

generally taken to indicate that the surface buoyancy forcing generates narrow updrafts flanked by broader downdrafts (Moeng and Rotunno 1990; Hogan et al. 2009; Ansmann et al. 2010), an observation supported in subsequent analyses. The skewness is greatest in the upper half of the CBL because of penetrative thermals in this region.

FIG. 3. Composite CBL evolution and surface forcing for 138 ShCu days. (a) Mean boundary layer height (black line), lifting condensation level (blue dashed line), vertical velocity variance (gray shading), vertical velocity skewness (red contours; contour interval = 0.1 starting at 0.05), cloud-base heights (circles), and cloud fraction (color fill in circles). The mean daily maximum CBL height is indicated. (b) The inner quartile ranges (IQRs) of sensible (red) and latent (green) heat fluxes from site E14. (c) The IQRs of CBL (black) and LCL (green) temporal evolution. (d) The IQR of cloud fraction.

The median time of cloud initiation (defined here as cloud fraction $> 5\%$) is 1015 CST, with cloud fraction then increasing to $\sim 10\%$ by the time the CBL top encroaches on the LCL at 1135 CST (Fig. 3a). The good agreement between CBL height, LCL height, cloud-base height, and cloud initiation time provides confidence in the measurement and analysis strategy. Following cloud initiation, cloud bases remains closely coupled to the CBL up until ~ 1500 CST then begin to decouple. Cloud fraction peaks at $\sim 23\%$ at 1455 CST then diminishes through the late afternoon.

2) STRATIFICATION

Figure 4a shows the composite potential temperature and Brunt–Väisälä buoyancy frequency $\{N = [(g/\theta)(\partial\theta/\partial z)]^{1/2}\}$ determined from the AERI on ShCu days. The composite indicates a surface-based nocturnal inversion in the morning giving way to a daytime superadiabatic layer overtopped with approximately neutral stratification within the CBL (note the superadiabatic portion is shown in white in Fig. 4a). The CBL top, determined from the DL, corresponds to a layer of increased stratification aloft. A maximum in N is observed in the afternoon (red contours for emphasis) somewhat above the

CBL top, which is consistent with the encroachment model of CBL growth into the background stratification, which results in a sharpening capping inversion aloft (Stull 1988). Starting at ~ 1800 CST, the surface stability increases because of negative sensible heat fluxes (see Fig. 3b). As a result, the CBL decouples, and a residual layer forms. The timing of this transition corresponds with the observed decoupling and dissipation of clouds shown in Fig. 3a.

Figure 4b compares the AERI-derived potential temperature profile with that from the 1130 CST radiosondes. Because of relatively coarse vertical resolution near the CBL top (~ 1 km; Turner and Löhnert 2014), it is apparent that the AERI does not resolve the capping inversion as well as the radiosondes. Rather, the AERI tends to overestimate the stratification in the upper CBL and underestimate the stratification in the capping layer. The correlation between the radiosonde and AERI-derived stratification is modest ($r^2 = 0.45$), and the slope is close to one to one (not shown). These characteristics are consistent with those found in Blumberg et al. (2017), who show that bulk convective indices (e.g., lifted index) compare well between the AERI and radiosondes, whereas integrative measures (i.e., CAPE) are in lesser agreement.

FIG. 4. Overview of temperature stratification on ShCu days. (a) Composite Brunt-Väisälä frequency (shaded; white indicates imaginary data and thus superadiabatic) and potential temperature (black contours; contour interval 1 K) from the AERI profiler. The mean CBL height from the Doppler lidar is shown as a solid black line. Red contours (contour interval = 0.0002 s⁻¹) have been added to emphasize the increased stratification above the CBL in the afternoon. Heights are normalized by the maximum daily CBL height. (b) Comparison of the mean potential temperature profiles determined by radiosondes (red) and AERI (black) at 1130 CST. Shown are the mean (solid) and one-standard-error ranges (dashed).

3) HUMIDITY

Figure 5 shows the clear-air relative humidity (Fig. 5a), mixing ratio (Fig. 5b), and the vertical gradient of the mixing ratio ($\partial q/\partial z$; dashed blue contours, Fig. 5b) during the composite ShCu day. RH increases along the top edge of the growing CBL through the morning, reaching a maximum of 71% at 1415 CST, which is ~ 40 min prior to the observed maximum in cloud fraction. The lowest CBL humidity occurs in the late afternoon near the surface.

Figure 5b shows that high mixing ratio air (~ 13 g kg⁻¹) expands upward in concert with the CBL growth because of both the redistribution of moist air initially confined near the surface and addition of water via latent heat fluxes. Like Fig. 5a, these data also indicate CBL drying in late afternoon, which presumably corresponds with entrainment of dry air from aloft (an additional analysis of advective fluxes showed little contribution in the mean). To this end, entrainment is related to the vertical gradient of the mixing ratio along the CBL top, which becomes increasingly negative as the CBL impinges on the drier free troposphere in the afternoon (Fig. 5b, contours). There is, for example, a peak in the magnitude of $\partial q/\partial z$ near the time of the maximum CBL height, which also closely corresponds to the increase in buoyancy frequency documented in Fig. 4.

b. ShCu versus clear-sky turbulent profiles

Here, we examine differences in CBL turbulence between clear-sky and ShCu conditions. Figure 6 presents

time-mean profiles of VV variance (Fig. 6a) and skewness (Fig. 6b) for the 138 ShCu days and the set of 108 clear-sky reference days. Only the hours of 1100–1700 CST are included in this analysis because the CBL height is relatively slowly evolving during that period. Unlike the composites above (i.e., Figs. 3–5), in these profiles, the height is normalized by the CBL height for each hour rather than the daytime maximum CBL height. The variance is then normalized by the square of the convective velocity:

$$w^* = \left(\frac{g}{\theta} \overline{w' T_v'} Z_i \right)^{1/3},$$

where Z_i , the CBL height, is determined from the lidar and the virtual heat flux ($\overline{w' T_v'}$) is from QCECOR data.

Despite the difference in sky cover, the clear-sky and ShCu CBL variance profiles are similar, both collapsing to a curve with peak normalized intensity of ~ 0.275 occurring at $\sim 0.3 Z_i$ (black and green curves in Fig. 6a). The differences between these curves are not statistically significant over most of the depth of the CBL (open circles indicate statistical similarity of the means). The similarity is not surprising since most hours on ShCu days have relatively low cloud fraction. We also note that our variance profiles are similar in form, though lower in magnitude, than those reported by Berg et al. (2017), who examined yearlong data from the same DL at ARM SGP. The reduction in magnitude is due to the

FIG. 5. Composite analysis of humidity as a function of normalized height and time of day from the Raman lidar. (a) RH and (b) mixing ratio (shaded) and the vertical gradient of the mixing ratio (blue contours; contour range: -3 to $-4.5 \text{ g kg}^{-1} \text{ km}^{-1}$; contour interval: $0.5 \text{ g kg}^{-1} \text{ km}^{-1}$). The dashed black line is the mean normalized CBL height, where the normalization is by the daily maximum CBL height (Z_i).

use of virtual temperature, rather than temperature, in the formulation of w^* , which tends to make w^* systematically larger (i.e., $T_v > T$).

We have also compared other aspects of CBL evolution between the clear-sky and ShCu datasets (analysis not shown). The CBL grows $\sim 350 \text{ m}$ deeper on ShCu days than on clear-sky days (2082 m compared with 1733 m , statistically significant at 95% confidence level). However, the mean sensible and latent fluxes are almost identical (and not statistically different) between these datasets. As such, the difference in CBL height is due to increased temperature stratification on clear-sky days, which slows

the CBL growth (not shown, statistically significant at 95% confidence interval). In addition, clear-sky days have lower humidity as expected (Zhang and Klein 2010) and thus much higher LCLs than ShCu days (maximum LCL 2546 m vs 2145 m , statistically significant at 95% confidence level). Combined, the suppressed CBL growth and higher LCL cause no clouds to form on the selected clear-sky days.

To further examine the turbulence dependence on cloudiness, the variance for hours with low ($0 < \text{CF} < 0.3$), medium ($0.3 \leq \text{CF} < 0.5$), and high ($\text{CF} \geq 0.5$) cloud fractions is examined and compared with the clear-sky variance profiles. For low cloud fraction, the

FIG. 6. Mean profiles of the higher-order moments of vertical velocity on ShCu and clear-sky days: (a) the normalized variance profiles, where height is normalized by CBL height and variance is normalized by the square of the convective velocity; (b) height-normalized profiles of vertical velocity skewness. In each panel, the solid green line represents the average clear-sky profiles. The black line indicates the average ShCu profile. The other colored lines indicate the variance and skewness for hours with low, middle, and high cloud fractions. Filled markers indicate data points that are significantly different than the clear-sky profile. Open markers indicate statistical insignificance.

FIG. 7. Cloud-base statistics for the 1578 individual clouds sampled. (a) Cloud duration above the lidar. (b) Cloud chord length. (c) Mean cloud-base updraft speed (threshold = 0.5 m s^{-1}). (d) The 95th percentile of cloud-base updrafts (threshold = 0.5 m s^{-1}).

variance is not statistically different than the clear-sky profile (cf. blue and green lines in Fig. 6a). In contrast, for medium cloud cover ($0.3 < \text{CF} < 0.5$), the variance is systematically higher over the depth of the CBL (red line, statistically significant at 95% confidence level, Fig. 6a), suggesting that in this class of sky cover, there are interactions and feedbacks between the clouds and the CBL turbulence. For example, it is possible that the more energetic convective updrafts and downdrafts (and thus higher variance) more readily overcome the energy barrier at the CBL top and thus initiate higher cloud fractions. However, periods of even higher cloud fractions ($\text{CF} > 0.5$; magenta line, Fig. 7a) do not correspond to a further increase in the variance but rather converge back to the clear-sky mean (i.e., the profile is not statically different). We note that these results differ from those in Chandra et al. (2010), wherein the normalized variance during ShCu conditions was shown to be lower than during clear-sky conditions. The source of these differences is not known, though differences in measurement technique (radar scattering from insects vs lidar scattering from aerosol) may be one factor.

The skewness profiles for both clear and cloudy days (Fig. 6b) indicate positive skewness throughout the CBL

and maximizing near $0.7Z_i$. This distribution is generally taken to represent narrow updrafts flanked by broad downdrafts (Moeng and Rotunno 1990; Hogan et al. 2009; Ansmann et al. 2010). Notably, the mean skewness on ShCu days is somewhat larger than on the clear-sky days (cf. red and green lines, Fig. 6b), consistent with the findings of LK15. The same is true, in general, of profiles for hours with low and medium cloud fraction. For cloud fraction above 50%, however, there is a marked reduction in skewness, suggesting that the updrafts and downdrafts during periods of higher cloud cover are closer in scale and intensity. This change in skewness may reflect a change in convective organization within the CBL that contributes to the development of more organized cloud cover, a feedback between cloudiness and CBL turbulence, or some combination thereof.

c. Subcloud circulations

1) CLOUD-BASE STATISTICS

In total, 1791 individual cumuli are identified in the 5-yr dataset. Among these clouds, mass flux and updraft data were available for subset of 1578 clouds (the excluded clouds had insufficient vertical velocity data).

FIG. 8. Overview of cloud properties based on mass flux. (a) Distribution of observed cloud-base mass flux. (b) PDFs of cloud chord length for positive (blue) and negative (brown) cloud-base mass flux. (c) Diurnal cycle of positive (red), negative (dark blue), and unknown (magenta) cloud occurrences. The percentage of positive cloud-base mass flux clouds to all clouds in a given hour is indicated.

The cloud-base statistics for these clouds are summarized in Fig. 7. The mean (median) cloud residence time over the lidar is 180 (125) s and the mean (median) chord length is 1185 (797) m, respectively. The cloud duration and chord length distributions are both positively skewed and therefore dominated by short duration and relatively small (sub-kilometer scale) clouds.

The distributions of the cloud-base mean and maximum updrafts are shown in Figs. 7c and 7d. The mean (median) updraft across all clouds is 0.89 (0.78) m s^{-1} , and the mean (median) of the 95th percentile updrafts, W_{95} , is 1.97 (1.77) m s^{-1} , though individual updraft speeds are as high as 7 m s^{-1} . The W_{95} updrafts are of particular interest because it is expected that they are most likely to survive into the core of the cloud. For this reason, we later characterize cloudy periods by these updrafts.

Figure 8a shows the distribution of cloud-base mass flux [see Eq. (1)]. Like the other statistics, this distribution is positively skewed such that 62.3% of clouds with reported mass flux values have a net positive mass flux and the remaining 37.7% possess negative mass flux. The fraction of all clouds (including clouds with no mass flux computed) possessing positive mass flux is $\sim 54\%$ (not shown). Collectively, these results imply that most of the observed clouds export mass (and other scalars) from the CBL into the cloud layer. As will be shown below (in Fig. 10), the minority of clouds with negative mass flux lack coherent subcloud forcing for their continued growth and are thus presumed to be decaying.

Notably, the cloud duration and chord length distributions differ between the positive and negative mass flux subsets, as summarized in Table 1. The median cloud duration for positive mass flux clouds is 137 s compared with 102 s for negative mass flux clouds. The corresponding median chord lengths are 855 versus 647 m, respectively. The population means, shown in Table 1, are significantly different at the 95% confidence interval. The PDFs of cloud chord length for negative

and positive mass flux clouds (Fig. 8b) show a clear shift toward smaller chord lengths for the negative mass flux clouds, though there are still many negative mass flux clouds at larger sizes. These findings are consistent with those of Rodts et al. (2003), where the smallest subset of clouds were observed to contribute negatively to the net mass flux along flight transects through a field of cumuli.

The diurnal cycle of the occurrence of positive mass flux clouds (red bars), negative mass flux clouds (blue bars), and unknown mass flux clouds (magenta bar) is shown in Fig. 8c. The fraction of positive mass flux clouds (relative to all clouds in a given hour) is greater than 50% for the first half of the day, peaking at 59% between 1100 and 1200 CST. This is consistent with the trend toward increasing cloudiness (i.e., a net production of clouds) up until the maximum in cloud fraction at 1455 CST (as shown in Fig. 1). After 1500 CST, positive mass flux clouds are in the minority, consistent with decreasing cloud cover and a preponderance of negative mass flux clouds.

The joint distribution of cloud size and W_{95} updrafts for the subset of positive mass flux clouds is examined in Fig. 9a. Also shown are the bin-mean W_{95} updraft values (white dots). There is an overall trend toward stronger updrafts for wider clouds up until a cloud chord length of ~ 2000 m. This size dependence is also apparent when the cloud chord length is normalized by the CBL depth and the W_{95} updrafts are normalized by the convective velocity (w^*) (Fig. 9b). To be specific, the normalized velocity increases from $\sim 0.6w^*$ for the smallest clouds to more than w^* for clouds widths of $\sim 0.5Z_i$ and then as cloud chord length approaches the CBL depth. Both Ansmann et al. (2010) and LK15 showed similar “increasing then asymptoting” relationships between lidar-derived updraft width and intensity at the top edge of the CBL during ShCu conditions. Ansmann et al. (2010) suggested that a second-order polynomial can be used to describe this relationship, which may be useful in

TABLE 1. Cloud-base statistics for individual shallow cumuli.

Cloud type	Cloud count	Mean duration (s)	Median duration (s)	Standard deviation (s)	Mean chord length (m)	Median chord length (m)	Standard deviation (m)	Mean updraft (m s^{-1})	Median updraft (m s^{-1})	Mean maximum updraft (m s^{-1})	Median maximum updraft (m s^{-1})
All clouds	1791	180	125	168	1185.5	796.97	1258	0.8958	0.78	1.97	1.77
Positive mass flux	998	194.84	136.97	181.71	1269.6	855.07	1328.3	1.15	0.973	2.566	2.38
Negative mass flux	573	150.61	102.36	146.09	1025.9	646.96	1206.7	0.62	0.443	0.94	0.822

parameterizing the size dependence of updraft strength (i.e., [Rochetin et al. 2014a,b](#)). To this end, we have fit a second-order polynomial to our data, the coefficients of which are shown in [Fig. 9b](#).

2) COMPOSITE SUBCLOUD CIRCULATION

The median vertical velocities throughout the subcloud layer for the positive and negative mass flux subsets of clouds are shown in [Figs. 10a and 10b](#). Also shown are the updraft and downdraft fractions for the positive and negative mass flux clouds, respectively ([Figs. 10c,d](#)). The composites are normalized by cloud-base height and cloud duration such that cloud base corresponds to a height of 1 and the cloud duration is from -0.5 to 0.5 . For reference, the mean CBL height is also shown, as is the distance between the CBL top and the cloud base. The positive mass flux clouds are clearly linked to a coherent subcloud circulation with an updraft that extends over the greater depth of the CBL ([Fig. 10a](#)). The updraft is slightly asymmetric in time with stronger updrafts preferentially occurring toward the leading cloud edge. The highest median updraft speeds ($\sim 0.8 \text{ m s}^{-1}$) are observed at 0.8 of the cloud-base height and the cloud base itself is, on average, $\sim 50 \text{ m}$ above the CBL top. Updrafts are present only $\sim 60\%$ of the time within the center of the updraft and along cloud base ([Fig. 10c](#)). That the updraft frequency is greatest along cloud base and in the upper CBL likely results from two factors. First, by design, our sampling focuses on the time clouds (and their associated updrafts) occur, thus favoring coherence aloft. Second, CBL updrafts typically grow upscale with height because of the merging of smaller rising plumes in the lower CBL ([Williams and Hacker 1992](#)).

For the positive mass flux clouds, the updraft is flanked by broad downdrafts that extend over the depth of the CBL. The median downdraft strength is $\sim -0.3 \text{ m s}^{-1}$, which is less than half the magnitude of the composite updraft centered under the cloud. This finding is consistent with the aforementioned skewness of vertical velocity on ShCu days ([Fig. 6](#)): narrow updrafts flanked by broad weak downdrafts. Interestingly, near the cloud edges, compact (i.e., short in duration) downdrafts are observed that are stronger than in the broad downdraft regions (especially near time -0.5 and height ~ 1). These compact downdrafts extend a short distance into the subcloud layer.

By comparison, the composite circulation beneath negative mass flux clouds ([Fig. 10b](#)) lacks a persistent updraft and is generally characterized by subsiding air along the cloud base; thus, supporting our interpretation of these clouds as either decaying or a portion of a cloud where the updraft did not pass directly over the lidar. Interestingly, despite the absence of an updraft, the

FIG. 9. Relationships between cloud size and the 95th-percentile (W_{95}) cloud-base updrafts. (a) Joint probability distribution of cloud chord length and W_{95} updrafts (shaded) and bin-mean updraft as a function of cloud size (white dots; bin size = 250 m). (b) Bin-averaged distribution of W_{95} normalized by the convective velocity (w/w^*) and cloud chord length normalized by CBL height (Z_i). The dots show the mean, and the bars indicate the standard error. Bins with <20 samples are indicated with open circles. A second-order polynomial fit is provided for the data, with coefficients indicated in the figure.

cloud-edge downdrafts remain intact and contribute to the net downward mass flux. These cloud-edge downdrafts are present 50%–60% of the time (Fig. 10c). While the PDF of cloud chord lengths for the negative mass flux clouds is shifted to smaller horizontal scales (as summarized in Table 1 and Fig. 8b), there are still many large negative mass flux clouds (>1-km chord length), such that it is unlikely that negative mass flux clouds are only an artifact of sampling “glancing” overpasses.

Rodts et al. (2003) suggest two mechanisms may contribute to cloud-edge downdrafts: (i) mechanical forcing from penetrative updrafts and (ii) evaporative cooling along the cloud edges. During aircraft penetrations of shallow cumuli, they found that the cloud-edge downdrafts were characterized by a sharp drop in virtual potential temperature and thus were likely driven by evaporative cooling, not mechanical forcing. Evaporative cooling was also found to generate “subsiding shells” around cumuli in large-eddy simulations (Heus and Jonker 2008). More recently, the three-dimensional characteristics of longwave radiative cooling have been implicated as another contributing source for these downdrafts (Klinger et al. 2017). To help identify the source of the cloud-edge downdrafts in our sample of cumuli, we conducted a second analysis using conditional sampling of updrafts in the upper CBL (at $0.95Z_i$) irrespective of their link to clouds (Fig. 11). The updrafts were then binned into cloudy and cloud-free samples based on the maximum lidar backscatter in the column

(using the cloud backscatter thresholds discussed above). In total, 1881 cloudy and 3343 clear updrafts were sampled [we note that there are slightly more cloudy updrafts sampled than were clouds (1791) because some continuous clouds have multiple, separate updrafts]. Based on Fig. 11, two key findings are apparent: (i) both cloud-free and cloudy updrafts are flanked by coherent downdrafts of comparable magnitude and that extend the same distance into the CBL, and (ii) the cloudy updrafts are stronger and wider than their clear-air counterparts (mean of ~ 1.11 vs ~ 0.83 m s⁻¹).

The first finding indicates that the downdrafts observed near the cloud edges are likely due to mechanical forcing, not evaporative or radiative cooling, since downdrafts are equally as strong between clear and cloudy samples. This contrasts with the conclusion of Rodts et al. (2003), who implicate evaporative cooling as the primary mechanism. Since most clouds in our sample are presumed to be “forced” cumuli, this result is not surprising. This result holds, however, even when considering only cloudy updrafts associated with high LWP clouds ($LWP > 50$ g m⁻²; sample size 607 cloudy updrafts), which might be more likely to produce evaporatively cooled shells (magenta line, Fig. 11c). We do note, however, that the downdrafts for these higher LWP clouds are somewhat narrower (with respect to updraft scale) than for the other samples. These results do not, however, fully explain the existence of cloud-edge downdrafts for negative mass flux clouds. These

FIG. 10. Composite subcloud vertical velocity analysis for (a) positive mass flux clouds and (b) negative mass flux clouds. The height is normalized by cloud-base height (Z_{cb}), and the time is normalized by cloud duration such that all clouds occur between -0.5 and 0.5 (vertical dashed lines). Time increases to the left. The mean cloud base is indicated in white circles. The mean distance between the cloud-base height and the CBL top (ΔH) is indicated, and the CBL top is shown as a dashed black line. The number of clouds in each sample (N) is indicated. (c) Updraft and (d) downdraft fraction for positive mass flux and negative mass flux clouds, respectively.

downdrafts may exist because of forcing by an updraft that was not sampled (e.g., only the cloud edge moved over the lidar), evaporative cooling, radiative cooling, or some combination thereof.

Unfortunately, the data currently available are insufficient to more fully probe the source of these

downdrafts. Originally, the Doppler lidar and Raman lidar were not collocated at the SGP site, which prevented assessment of the moisture and temperature anomalies in the downdraft regions. This problem was rectified in November 2015 when the Raman lidar was moved to be within 10 m of the Doppler lidar, opening

FIG. 11. Composite vertical velocity analysis of updrafts conditionally sampled at $0.95Z_i$. Heights are normalized by Z_i , and time is normalized by updraft duration. (a) Composite vertical velocity of all clear updrafts. (b) Composite vertical velocity of all cloudy updrafts. (c) Mean vertical velocity at $0.95Z_i$ for clear (blue), cloudy (red), and high-LWP ($LWP > 50 \text{ g m}^{-2}$) clouds (magenta; sample size is 607 clouds). The filled markers indicate locations where the means are statistically different at the 95% confidence level.

the possibility of future observational study of these (and other) cloud and boundary layer processes.

The second finding is also interesting and consistent with other investigations (e.g., LK15): cloudy updrafts are stronger than their cloud-free counterparts. This is especially the case for updrafts linked to clouds with higher LWP ($LWP > 50 \text{ g m}^{-2}$; magenta line, Fig. 11c), which have a peak updraft strength of 1.35 m s^{-1} . Furthermore, the stronger–cloudy updrafts tend to be broader, with the mean cloudy updraft $\sim 164 \text{ m}$ wider than clear updrafts. In fact, the PDF of chord lengths for cloudy updrafts is systematically shifted toward larger values compared to the clear updrafts (supplemental Fig. S3). All of the above differences in means are statistically significant at the 95% confidence level.

Finally, we also examined composite subcloud circulations based on terciles of the cloud chord length ($L < 507 \text{ m}$, $507 \leq L < 1167 \text{ m}$, $L \geq 1167 \text{ m}$), rather than the sign of the mass flux (supplemental Fig. S2). Each bin contained ~ 595 cloud samples. Despite the differences in cloud size, the same general characteristics for subcloud circulations were found across size bins: a coherent updraft centered on the time of the cloud overpass flanked by weaker downdrafts, including stronger downdrafts adjacent to the cloud edge in the upper CBL. The updraft intensity (and frequency of occurrence) was greatest for clouds with horizontal scales between 0.5 and 1 km and weakest for the smallest subset of clouds.

3) ENERGY BARRIER

To contextualize the observed cloud-base updrafts, it is necessary to quantify the strength of the “energy barrier” separating the CBL from the cloud layer. To do so, the CIN to reach 300 m above the LCL is computed based on mixed-layer parcels (lowest 50 hPa) for the daily 1130 CST radiosonde. The 300-m displacement above the LCL follows from ZK13, wherein “active” cumulus clouds were classified as those exceeding a depth of 300 m. We note that this definition of CIN is different than the standard CIN defined for lifting a parcel to the level of free convection, although ZK13 found that clouds with depths greater than 300 m had in the vast majority of cases reached their levels of free convection (Fig. 2 of ZK13). Since the 1130 CST radiosondes tend to be near the initiation time for ShCu (ZK13; see also Fig. 3a), these data are expected to be representative of the ambient conditions affecting cloud growth early in the day. However, these CIN data do not capture diurnal changes in energy barrier strength associated with changes in the CBL depth later in the afternoon.

To overcome this limitation, we use the AERI profile data to examine the temporal evolution of stratification near cloud base. Specifically, we measure the mean

buoyancy frequency (N) in the layer of depth H between the CBL top (Z_i) and 300 m above the LCL. The distance (H) describes an average parcel displacement necessary to initiate active convection at each point in time. The stratification (N) and displacement (H) are then combined to form a single energy barrier variable (NH), which has units of meters per second and is thus directly comparable with updraft velocities at cloud base. Unfortunately, AERI data are available for only a limited set of 91 ShCu days. An example of the NH analysis for one ShCu day is shown in Figs. 12a and 12b.

Figures 12c and 12d shows the climatological distributions of N , H , and NH along with the inner quartile range of $(\text{CIN})^{1/2}$ at 1130 CST. At that time, the $(\text{CIN})^{1/2}$, which also has units of meters per second, and the NH data indicate a similar range. More generally, NH indicates high-energy barrier values in the morning hours ($\sim 9 \text{ m s}^{-1}$) that decay to a nearly consistent value ($\sim 4 \text{ m s}^{-1}$) in the afternoon. Examination of Fig. 12c indicates that the daytime variation in NH is dominated by variations in the distance between the CBL top and the LCL (i.e., H). To be specific, N is relatively constant through the afternoon, whereas H reaches a minimum near the time of maximum cloud fraction (1455 CST). After the evening transition begins, the values of NH again increase.

4) UPDRAFT FRACTION AND CIN-BASED CLOSURE

The functional relationships among vertical velocity, energy barrier, and cloud development are examined by binning observations into 30-min periods. Each period is characterized in terms of a total updraft fraction (i.e., fraction of the 30-min period with cloud-base updrafts $> 0.1 \text{ m s}^{-1}$), mean energy barrier value, and maximum cloud-base updraft. In so doing, a total of 2070 half-hour periods are considered, each possessing a distribution of CBL and cloud-layer properties (though many hours have no clouds). From these aggregated data, we test the hypothesis that dimensionless cloud inhibition (CI) parameters can be used to explain the variation in updraft fraction and mass flux.

The two dimensionless parameters considered are

$$\text{Cloud Inhibition 1} = \frac{\sqrt{\text{CIN}}}{W_{\text{cb}}} \quad \text{and}$$

$$\text{Cloud Inhibition 2} = \frac{NH}{W_{\text{cb}}},$$

where W_{cb} is the representative updraft velocity, taken here to be the maximum W_{95} updraft across the sample of clouds occurring within a 30-min window. In general, it is expected that when the strongest updrafts (denominator) become large relative to the energy barrier

FIG. 12. Energy barrier variables. (a) Example of the stratification (N^2) from the AERI profiler, CBL height (Z_i , black), LCL (red), and LCL + 300 m (blue) on a ShCu day. (b) Time evolution of NH corresponding to (a). (c) Climatology of N (red shaded IQR) and H (transparent blue IQR) on ShCu days. (d) Climatology of the energy barrier strength (NH; blue shaded IQR) on ShCu days. Also shown in (d) is the IQR and median value of the square root of CIN at 1130 CST from the radiosonde data for all ShCu days with radiosonde data (red bar, black line).

term (numerator), the updraft fraction, and thus updraft mass flux, will increase (Bretherton et al. 2004; Park and Bretherton 2009; Fletcher and Bretherton 2010).

The observed relationships between the 30-min updraft fraction and the CI parameters are shown in Fig. 13. Despite scatter in the data, both analyses reveal the same functional relationship: the updraft fraction sharply increases with decreasing CI. This is especially clear when considering the bin-median values of updraft fraction (CI bins of 0.5), shown in each figure panel as a solid black line. This finding is generally consistent with the proposed exponential form of the closure for updraft fraction employed in the CIN-based ShCu parameterization detailed in Park and Bretherton (2009) and applied in Fletcher and Bretherton (2010) and the complimentary error function formulation in Bretherton et al. (2004). For example, also shown in each panel is a set of complimentary error functions of the form

$$a = 0.5 \operatorname{erfc}(c\text{CI}),$$

where a is the updraft fraction (percent of time with $w > 0.1 \text{ m s}^{-1}$), c is a constant (arbitrarily shown here for $c = 0.6, 0.8$, and 1), and CI is either of the cloud inhibition parameters. Such results have not been previously demonstrated from observations alone.

The differences between the two CI parameters are relatively subtle, suggesting that the 1130 CST radiosonde data coupled with updraft observations are often

sufficient to capture the overall form of this relationship. On the other hand, the somewhat lesser scatter in the NH-based CI analysis likely reflects a more accurate accounting of the diurnal variations in energy barrier relative to updraft strength. Mesoscale variability (e.g., convective rolls) that is not fully captured by observations of a single vertical column of the atmosphere (cf. Berg and Stull 2002) is one likely source of scatter in both analyses.

It is also interesting to note that the maximum cloud LWP for each period, as determined from the MWR, varies with CI (color fill on data points in Fig. 13). Since LWP is a crude proxy for cloud depth (assuming liquid water content is a fixed fraction of the adiabatic value and similar cloud-base temperatures across all clouds), this suggests that the CI parameters capture some of the dependence of cloud vertical extent on the CBL forcing: deeper clouds become more prevalent as the strength of updrafts become large relative the strength of the energy barrier. This finding is consistent with modeling results showing that CIN-based closure applies even across the transition to deeper convection (Fletcher and Bretherton 2010; Hohenegger and Bretherton 2011).

The CI parameters (in either formulation) also provide a metric for grouping days with different CBL and cloud-layer characteristics. For example, ShCu days can be characterized by the median CI value for the hours 1130–1500 UTC, corresponding with the midday period

FIG. 13. Relationship between the half-hourly updraft fraction (ordinate) and the two dimensionless CI parameters (abscissa). (a) CI determined as the $(\text{CIN})^{1/2}/W_{cb}$ and (b) CI determined as NH/W_{cb} . In both panels, the dots represent the fraction of each 30-min period characterized by updrafts ($w > 0.1 \text{ m s}^{-1}$), and the color fill is the maximum LWP for that period. The solid black line is the bin-median value, using bin sizes of 0.5 CI. There are fewer available days with valid NH/W_{cb} data, thus, the more limited number of points. Also shown in each panel are a family of complimentary error functions (see text).

where the CBL is close to stationary and exhibits higher cloud fraction. Days can subsequently be grouped by CI terciles, wherein it is expected that higher terciles correspond to conditions where it is increasingly difficult to initiate clouds.

To this end, Figs. 14a–c show the CBL composite evolution for each CI tercile using the same conventions as in Fig. 3. The differences in diurnal evolution in LWP, cloud fraction, and the distance between the LCL and the CBL top are also examined (Figs. 14d–f). Taken together, these composite data indicate clear differences across the terciles consistent with the expected result. For example, the time of cloud initiation is progressively later and the cloud fraction lower for increasing CI (note that not all hours have statistically different means). The mean LWP also varies across the terciles, with higher LWP, and presumably deeper clouds, for low CI. While the CBL height variation with CI is minimal (not statistically significant), there is an increased separation between the CBL top and LCL across the terciles (Fig. 14f). Specifically, the upper terciles have a larger separation during midmorning and thus larger energy barriers that must be overcome to initiate clouds.

Interestingly, the differences in sensible and latent heat fluxes between the terciles are not statistically significant (not shown). This finding suggests the role of relatively humidity in modulating cloudiness, as found by ZK13 and LK15. This inference is confirmed by inspecting the corresponding tercile composites of the relative humidity (Fig. 15). The relative humidity is higher throughout the lower troposphere (normalized height 0–1) on days with high cloud fraction and early cloud onset (Fig. 15a) and diminishes with increasing CI (Figs. 15b,c). These

differences are somewhat easier to interpret from the mean morning (0600–1200 CST) profiles, which indicate RH of $\sim 70\%$ in the upper boundary layer for low CI, as compared to $\sim 62\%$ for the second tercile composite. While this RH dependence is not a new result, it demonstrates the consistency of the CI analysis with ZK13 and LK15.

4. Discussion

While the CIN-based closure results are encouraging, in order for such a closure to be useful, the cloud-base updrafts must be predictable from the boundary layer turbulent kinetic energy (TKE) or surface forcing (w^*). Fletcher and Bretherton (2010), for example, parameterize the cloud-base updraft speed in terms of the CBL TKE as

$$W_{cb} = a\text{TKE}^{1/2} + b,$$

where a and b are empirical constants. One might expect a similar functional relationship between cloud-base updrafts and the observed VV variance, which is the vertical component of TKE. Figure 16, however, shows that there is almost no predictive skill in relating CBL VV variance, or even CBL updraft strength (W_{cbl} , defined as the maximum vertical velocity in the CBL below $0.66Z_i$), to cloud-base updrafts and thus cloud-base mass flux. While the CBL variance does (modestly) predict the maximum CBL updrafts ($r^2 = 0.621$; Fig. 16a), it fails to provide a meaningful estimate of updrafts observed at cloud base ($r^2 = 0.029$; Fig. 16b). In fact, even the maximum CBL updrafts for each 30-min period do not adequately predict the maximum cloud-base updraft strength

FIG. 14. Composite CBL evolution stratified by terciles of the daily mean CIN-based CI parameter. (a)–(c) CBL variables as in Fig. 3 and including maximum CBL depth (Z_i), number of days in the sample (N), and the maximum LWP (LWP_x); the colors from the dark red at the top of the legend to the bottom dark blue represent 35% and 5%, respectively, with a color interval of 5%. The CI tercile ranges are $CI \leq 1.64$, $1.64 < CI \leq 2.67$, and $CI > 2.67$ and are indicated above (a)–(c). Also shown are diurnal cycles of (d) LWP, (e) cloud fraction, and (f) the difference between the CBL top and the LCL height (δH) normalized by Z_i . Red, blue, and green lines indicate terciles 1–3, respectively. Standard errors are shown, and means that are statistically different at the 90% confidence interval are shown in the filled markers.

($r^2 = 0.075$; Fig. 16c). These results are not particularly surprising in that the CBL updrafts must work against buoyancy to penetrate to cloud base, which is typically slightly above the CBL top. It is also consistent with the analysis in Fig. 9b, where we showed that the cloud-base updrafts, relative to w^* , vary with the normalized updraft width. To this end, other recent works indicate that CBL updraft width is an important factor in updraft strength and that the skewness of the velocity field is related to the extent of penetrative thermals (Rochetin et al. 2014a,b; LK15). This dependence of cloud-base updrafts on normalized updraft width also limits the applicability of the mass flux parameterization proposed by Grant (2001), wherein the mass flux was assumed to be linearly proportional to the convective velocity. With this in mind, future work should target better understanding the factors controlling the covariation of updraft width and strength within the CBL and the selection of updrafts that go on to form clouds.

5. Summary and conclusions

This study examines the boundary layer controls on ShCu development at ARM SGP using a suite of collocated remote and in situ sensors. In the first portion of

the paper, the composite structure and evolution of the ShCu-topped CBL is documented. As part of that analysis, the differences between clear-sky and ShCu turbulent profiles are examined, revealing increasing skewness of the vertical velocity for cloudy periods (with cloud fraction < 0.5) and increased vertical velocity variance for periods with cloud fraction between 0.3 and 0.5. These results differ from those of Chandra et al. (2010), where it was found that cloudy periods had systematically lower variance and lower skewness than the clear-sky conditions. These differences warrant further investigation.

In the second part of the paper, the properties of individual cumulus clouds are examined, including the structure of the subcloud circulation. It is found that $\sim 63\%$ of the sampled clouds exhibit positive mass flux. For these clouds, a coherent subcloud updraft extends through the depth of the CBL and is flanked by broad, weak downdrafts. In contrast, for negative mass flux clouds, the coherent updraft is absent. Interestingly, both sets of clouds exhibit compact downdrafts adjacent to the cloud edge that penetrate some distance into the subcloud layer. These downdrafts are also present on dry thermals in the upper CBL and thus are likely the result of mechanical processes. This finding adds to the

FIG. 15. (a)–(c) Composite analysis of CBL RH for terciles of the CIN-based CI parameter. CI increases from (a) to (c); values as in Fig. 14. The CBL height for each group is shown by the dashed black line. (d) Average RH profiles for CI terciles [T1 (red), T2 (blue), and T3 (green)] for 0600–1200 CST, including standard errors. Thicker markers indicate means that are statistically different at the 10% confidence level.

ongoing dialogue concerning the origin of cloud-edge downdrafts in convective clouds, which have been attributed to evaporative cooling (Rodts et al. 2003; Hues and Jonker 2008), mechanical forcing (Jonas 1990; Klaassen and Clark 1985), and longwave radiative cooling (Klinger et al. 2017).

It was also found that cloud-base updraft strength is proportional to the updraft width for positive mass flux clouds. Specifically, the updraft speed (normalized by the convective velocity) increases with updraft width up until cloud width is comparable to CBL. This relationship is consistent with the findings of LK15 and

FIG. 16. Relationships among CBL and cloud-base vertical velocities. (a) CBL standard deviation of the vertical velocity and CBL maximum (or 95th percentile) updrafts (w_{cbl}). (b) CBL standard deviation of the vertical velocity and cloud-base updrafts. (c) CBL updrafts and cloud-base updrafts. The correlation coefficients shown in (a)–(c) are: 0.621, 0.029, and 0.075; the best-fit linear regression is also shown in each panel.

Ansmann et al. (2010) and underscores the importance of considering both updraft width and strength in parameterizing convective cloud processes (e.g., Rochetin et al. 2014a,b).

The cloud-base updraft data are subsequently used in conjunction with radiosonde and AERI thermal stratification data to examine the applicability of CIN-based mass flux closure assumptions. We find that the half-hourly updraft fraction varies strongly and inversely with the ratio of the energy barrier and updraft strengths. We name this dimensionless number the “cloud inhibition” (CI) parameter. This result provides observational support for the CIN-based closure assumptions that couple boundary layer and cloud-layer processes in some convective parameterizations (e.g., Bretherton et al. 2004; Park and Bretherton 2009).

We further demonstrated that CI distinguishes between different types of ShCu conditions. Low CI values (for the midday period) correspond to days with earlier cloud onset and higher cloud fractions, whereas high CI values indicate conditions where triggering ShCu becomes increasingly difficult. Days belonging to the former categorization show higher RH and also greater cloud LWP, which is a proxy for cloud depth. The variation in RH modulates, in part, the distance between the CBL top and the LCL such that it is more difficult for CBL updrafts to initiate clouds on high CI days.

A clear next step in this research is to extend these analyses to examine the transition from shallow to deep cumulus convection (Zhang and Klein 2010), as suggested by simulated results (Kuang and Bretherton 2006; Fletcher and Bretherton 2010). Such analyses could leverage the collocation of the Doppler lidars and ARM radars [e.g., Ka-band ARM zenith radar (KAZR)] to better quantify cloud depth and links between subcloud- and cloud-layer circulations.

Finally, it is also apparent that more research is required to improve the prediction of cloud-base updrafts based on CBL properties. Our analyses show that there is no clear link between the CBL vertical velocity variance and cloud-base updraft strength. Future observational analyses should more completely examine the relationships among updraft width, vertical coherence, moisture anomalies, and the strength of cloud-base fluxes.

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