Influences of drizzle on stratocumulus cloudiness and organization

Xiaoli Zhou¹, Thijs Heus², and Pavlos Kollias³,⁴

¹Department of Atmospheric and Oceanic Sciences, McGill University, Montreal, Quebec, Canada, ²Department of Physics, Cleveland State University, Cleveland, Ohio, USA, ³School of Marine and Atmospheric Sciences, State University of New York at Stony Brook, Stony Brook, New York, USA, ⁴Department of Environmental and Climate Sciences, Brookhaven National Laboratory, Upton, New York, USA

Abstract

Large-eddy simulations are used to study the influence of drizzle on stratocumulus organization, based on measurements made as part of the Second Dynamics and Chemistry of the Marine Stratocumulus field study-II. Cloud droplet number concentration (Nc) is prescribed and considered as the proxy for different aerosol loadings. Our study shows that the amount of cloudiness does not decrease linearly with precipitation rate. An Nc threshold is observed below which the removal of cloud water via precipitation efficiently reduces cloud depth, allowing evaporation to become efficient and quickly remove the remaining thin clouds, facilitating a fast transition from closed cells to open cells. Using Fourier analysis, stratocumulus length scales are found to increase with drizzle rates. Raindrop evaporation below 300 m lowers the cloud bases and amplifies moisture variances in the subcloud layer, while it does not alter the horizontal scales in the cloud layer, suggesting that moist cold pool dynamic forcings are not essential for mesoscale organization of stratocumulus. The cloud scales are greatly increased when the boundary layer is too deep to maintain well mixed.

1. Introduction

Stratocumulus clouds play a major role in Earth’s radiation budget through their high coverage over the oceans and hence large shortwave radiation albedo [Wood, 2012]. It has long been observed that stratocumulus are organized on mesoscale from hundreds of meters to hundreds of kilometers, most commonly occurring as closed cellular (dim walls, bright centers) or open cellular structures (bright walls, dim centers) [e.g., Agee et al., 1973; Atkinson and Zhang, 1996; Stevens et al., 2005b; VanZanten and Stevens, 2005; Wood and Hartmann, 2006; Sharon et al., 2006]. These 2–200 km wavelength mesoscale features (“meso β and meso γ” according to Orlanski [1975]) are found to dramatically affect cloud albedo [Rossow et al., 2002; Muhlbauer et al., 2014]. Organization due to the change of cloud condensation nuclei (CCN) can cause significant changes in the amount and distribution of cloud water, thus, exerting noticeable impacts on albedo. This is referred to as the “hydrological cycle effect” [Pincus and Baker, 1994] and is considered to be no less important than “Twomey effect” [Savic-Jovcic and Stevens, 2008, hereafter SS08], which is referred to as the reduction of albedo in the presence of fewer drops [Twomey, 1977]. Currently, the radiative response of stratocumulus to the changing climate constitutes a major uncertainty in general circulation models (GCMs), resulting from the unsatisfactory parameterization of the fraction and the spatial variability of low clouds [e.g., Bony and Dufresne, 2005; Bony et al., 2006]. A better understanding of mesoscale organization of stratocumulus is therefore important to constrain the cloud cover in GCMs.

The open cellular clouds embedded in stratiform cloud fields are referred to as pockets of open cells (POCs) and were frequently observed in recent Second Dynamics and Chemistry of Marine Stratocumulus (DYCOMS-II) field study [e.g., Stevens et al., 2003, 2005b]. POCs share a relatively similar large-scale environment with neighboring regions of more stratiform cloudiness [Stevens et al., 2003; Wood and Hartmann, 2006] and are often associated with locally enhanced drizzle [Stevens et al., 2005b, Comstock et al., 2005, 2007; Sharon et al., 2006; Wood et al., 2011]. Recently, Yamaguchi and Feingold [2015] showed that the size and the spatial distribution of precipitating elements serve to make a difference for cloudiness. Ovchinnikov et al. [2013] pointed out that the transition from closed to open cells occurs when the rain initiation time scale is shorter than the updraft time scale, which can be achieved by increasing cloud thickness or decreasing droplet number concentration and updraft velocity. We note, however, that their analyses did not consider the spatial variability of clouds, yet it has been found in both observational and numerical studies that drizzle leads to...
a positive liquid water path (LWP) skewness [Stevens et al., 1998; Wood, 2005; Yamaguchi and Feingold, 2015; Wood and Hartmann, 2006]. Hence, it is intriguing to see how such variability contributes to the transition from closed to open cells. Here we use a Large eddy simulation (LES) model and apply water budget analyses to shed some light on this question.

Other than the transition of cloud cover from closed to open cells, drizzle also regulates their mesoscale organization. Numerical studies have suggested that the dynamic response of subcloud moist cold pools, resulting from raindrop evaporation, plays a critical role in the formation and evolution of open cells [Wang and Feingold, 2009; SS08]. The boundaries of cold pools from adjacent cells collide and form surface convergence zone and new cloud formation, resulting in an oscillation of clouds and precipitation in location and strength [Feingold et al., 2010]. These studies suggest that the subcloud circulation, caused by cold pool, is critical in shaping the organizational structure of open cells. Yet no definitive study on the role of cold pool dynamics in regulating horizontal cloud scales has appeared in the literature. Here we further investigate the impact of moist cold pool dynamics on stratocumulus organization, with particular focus on cloud scales. Our study also briefly discusses the role of marine boundary layer (MBL) depth on cloud organization. This issue has attracted considerable interest since a recent observational study has found a strong correlation between cloud scales and boundary layer depth over the eastern subtropical oceans [Wood and Hartmann, 2006].

The remainder of this manuscript is organized as follows. Section 2 outlines the model setup and case description. Section 3 shows domain-averaged cloud albedo and the evolution of the flow. In section 4 we investigate the contribution of thermodynamic variability in the lowest 300 m above the surface to stratocumulus organization. The influence of drizzle to the transition from closed to open cells is discussed in section 5. Section 6 discussed the mean cloud properties on insufficient scales. In section 7 we discuss our results within the framework of previous work and summarize our findings.

2. Model and Case Description

In this study, the University of California, Los Angeles LES (UCLALES) is applied, which is well documented in Stevens et al. [1999, 2005a]. This model solves the Ogura-Philips anelastic equations and uses a third-order Runge-Kutta time integration. The advection of momentum is implemented using fourth-order central differences, and low-order upwind schemes are applied for scalars whenever the local gradient is too steep. The Smagorinsky-Lilly model is used to parameterize subgrid fluxes, and the microphysical processes are parameterized based on a simple two-moment bulk warm rain scheme following Seifert and Beheng [2001, 2006]. The parameterization of autoconversion, accretion, and self-collection of drops can be found in Appendix A of Seifert and Beheng [2001]. In this study, the cloud droplet number concentration is held constant. While this simplified treatment cannot address the detailed complexities of cloud-aerosol interactions and drizzle formation, the fundamental dynamics are a reasonable starting point for analyzing the impact of drizzle on the surrounding flow and its mesoscale influences. Total water mixing ratio (qt) is prognosed, from which water vapor mixing ratio (qv) and cloud water mixing ratio (qc) are diagnosed by saturation adjustment. Additional prognostic equations for rain water mixing ratio (qr) and raindrop number concentration (Nd) are solved.

The configuration of the simulations mainly follows the GEWEX Cloud System Study DYCOMS-II second research flight (RF02) model intercomparison case of Ackerman et al. [2009]. The only forcings that depart from Ackerman et al. [2009] are the surface boundary conditions. In this study, simulations are performed for a relatively long time and therefore interactive surface fluxes are applied after 2 h of spin-up. The lower boundary has the characteristics of water held at a fixed temperature of 292 K as is measured during DYCOMS-II [Stevens et al., 2003]. The surface fluxes are derived on the basis of similarity theory to get realistic feedbacks from the lower boundary, where surface roughness length is fixed at 0.1 mm which is calculated from the Charnock’s relation for friction velocity $u_\tau = 0.25 \text{ m s}^{-1}$. Note that the stratocumulus cases discussed here are representative for nocturnal cloud decks; no solar radiative absorption is included. This simplification is based on observations that drizzle tends to peak during the later part of the night when stratocumulus clouds are at their thickest [Comstock et al., 2004] and that the solar absorption-driven decoupling does not control the transition of closed cells to open cells [Wood et al., 2008].
The cloud droplet sedimentation is included in our model due to its importance for cloud top entrainment [Ackerman et al., 2004; Bretherton et al., 2007]. The sedimentation flux is parameterized following Ackerman et al. [2009] by assuming Stokes’ law (quadratic dependence of fall speed on size) for the fall speed of a cloud droplet and a lognormal distribution of droplet radii with a geometric standard deviation \( \sigma_g \)

\[
F = c [3/(4\pi \rho_l N_c)]^{2/3} (\rho_l c)^{5/3} \exp(4.5 \log^2 \sigma_g)
\]

where \( c = 1.19 \times 10^8 \) m \(^{-1} \) s \(^{-1} \) and \( N_c \) is cloud droplet number concentration and \( \rho_l \) and \( \rho_i \) are the density of air and liquid water. \( \sigma_g \), the geometric deviation of cloud droplet size distribution is set to 1.2 to be consistent with the DYCOMS-II aircraft-observed stratocumulus cloud drop size distribution for RF02 [vanZanten et al., 2005].

Three classes of simulations are considered. First, a suite of simulations with cloud droplet number concentration \( N_c = 5, 10, 15, 25, 200 \) mg \(^{-1} \) (hereafter NC5, NC10, NC15, NC25, and NC200) was performed on a 25.6 km \( \times \) 25.6 km domain. One additional NC10 simulation was performed on a 51.2 km \( \times \) 51.2 km domain, to investigate if the chosen domain was sufficiently large [de Roode et al., 2004, hereafter dR04; Seifert et al., 2015]. No appreciable differences were found between the large domain simulation and our standard simulations; thus, the remaining of the presented work focuses on the 25.6 km \( \times \) 25.6 km domain simulations. The upper limit of \( N_c = 200 \) mg \(^{-1} \) is chosen to represent nondrizzling stratocumulus conditions, while the remaining cases are associated with drizzle. The cloud droplet number concentrations of drizzling stratocumulus are chosen based on their ability to produce drizzle and are in general lower than the number of CCN to the extent that not all aerosols are activated. The lower \( N_c \) of 5 mg \(^{-1} \) and 10 mg \(^{-1} \) has been frequently observed in open cells and are not considered as extreme cases [Wood et al., 2011; Terai et al., 2014].

The second set of simulations consists of model perturbation to isolate the impact of subcloud drizzle evaporation on stratocumulus organization and albedo. First, an additional NC10 simulation omits raindrop evaporation below 300 m (hereafter NC10-NE, where NE is short for “No Evaporation below 300 m”). Second, a filtering procedure [Jonker et al., 2006] was applied to some simulations to evaluate the impact of the subcloud moisture and temperature variance on cloud organization. This filtering procedure spectrally removes scales larger than 200 m from the total water mixing ratio \( q_t \) and liquid water potential temperature \( \theta_l \) for all levels below 300 m above the sea surface. Finally, a set of simulations with different domain sizes \( L \) \( (L = 1.6 \) km, \( 3.2 \) km, \( 6.4 \) km, and \( 25.6 \) km) were conducted. These simulations allow us to understand the response of stratocumulus organization in the absence of sufficient length scales.

For all simulations, the domain top is 3 km, the horizontal grid spacing is \( \Delta x = \Delta y = 50 \) m, and the vertical mesh is stretched with \( \Delta z = 5 \) m at the surface and within 125 m of the initial inversion, while dilating following a \( \sin^2 \) law in the interior of and above the boundary layer following SS08. All cases are simulated for 15 h.

### 3. Domain-Averaged Cloud Albedo and Evolution of the Flow

First, the influence of drizzle on cloud albedo is investigated. The development of drizzle in stratocumulus leads to considerable changes in the cloud amount and variability. Although our simulations do not comprise shortwave radiative processes, we can quantify the impact of drizzle formation on clouds through the pseudo albedo [Zhang et al., 2005]

\[
A = \frac{\tau}{6.8 + \tau}
\]

where \( \tau = 0.19 \) LWP \( 5/6 N_c^{1/3} \) is the cloud optical depth and LWP is the liquid water path (vertical integral of liquid water). The LWP and \( N_c \) are in the units of kg m \(^{-2} \) and kg m \(^{-1} \), respectively. Hereafter, we will use the term albedo for brevity.

Table 1 lists the simulated cloud cover, LWP in the cloudy grids (in-cloud LWP), and domain-averaged albedo \( (A) \) at 5 h for different \( N_c \). The relative differences from NC200 defined as \( (X_{200} - X_{N_c})/X_{200} \) are shown in the parenthesis of Table 1, with \( X_{200} \) being variables of NC200 and \( X_{N_c} \) of the drizzling cases (i.e., NC10, NC15, and NC25). The relative differences in albedo have two components: one Twomey effect that is directly due to the change in \( N_c \), and one dynamic effect that is due to changes in LWP. The latter can be estimated by calculating the relative difference in \( A \) with the LWP of each simulation and an \( N_c \) of 200 mg \(^{-1} \).
Table 1 shows that the albedo of NC25 changes by 26.4% from NC200, while only 3.8% is contributed by changes in LWP, suggesting that the Twomey effect explains over 85% of the albedo gap between the two simulations. However, the contribution of Twomey effect reduces greatly with the increase of drizzle, e.g., it explains ~40% of the reduction in the albedo difference between NC10 and NC200 and ~20% between NC5 and NC200. If we calculate the albedo using the mean cloudiness and LWP, we find an increase in albedo of ~0.03 (Table 1), which is ~4% of the albedo of NC200 but ~30% of NC5, suggesting that the spatial variability of clouds has a significant contribution to the albedo of drizzling stratocumulus.

Figure 1 shows the temporal evolution of domain-averaged surface precipitation rate ($R$), cloud cover (cloud water mixing ratio $q_c > 0.01$ g kg$^{-1}$), inversion height (height of maximum $\theta_l$ gradient), in-cloud LWP (LWP in the presence of clouds), and albedo for different $N_c$ cases. Our simulations span the range from nondrizzling (NC200) to heavily drizzling cases ($R > 1$ mm d$^{-1}$, NC5; Figure 1a). As seen in Figures 1b and 1d, higher rain rates correlate with lower cloud cover and less in-cloud LWP. Near the end of the simulations, the reduction of LWP prevents the clouds from precipitating efficiently, and therefore the precipitation rates converge (Figure 1a). For nondrizzling clouds (NC200), the high in-cloud LWP and overcast cloudiness maintain for 8–10 h, after which the LWP decreases by almost a factor of 2 and the cloud cover falls to below 100% (Figures 1b and 1d). The abrupt drop of cloud water results from the deepening of the marine boundary layer (MBL; Figure 1c) through strong entrainment [Albrecht et al., 1995; Wood and Bretherton, 2004] such that the cloud top cooling is not able to

![Table 1. Domain Averages of Cloud Cover, In-Cloud LWP, and Pseudo albedo for NC10, NC15, NC25, and NC200 at 5 h.](image)

Table 1. Domain Averages of Cloud Cover, In-Cloud LWP, and Pseudo albedo for NC10, NC15, NC25, and NC200 at 5 h.

<table>
<thead>
<tr>
<th></th>
<th>NC200</th>
<th>NC25</th>
<th>NC15</th>
<th>NC10</th>
<th>NC5</th>
</tr>
</thead>
<tbody>
<tr>
<td>Cloud cover (%)</td>
<td>99.7</td>
<td>98.8 (1.1%)</td>
<td>95.3 (4.4%)</td>
<td>84.7 (15.0%)</td>
<td>58.9 (40.9%)</td>
</tr>
<tr>
<td>In-cloud LWP (g m$^{-2}$)</td>
<td>97.7</td>
<td>94.8 (9.9%)</td>
<td>65.2 (33.8%)</td>
<td>47.1 (52.5%)</td>
<td>30.0 (69.9%)</td>
</tr>
<tr>
<td>A (%)</td>
<td>67.9</td>
<td>50.0 (26.4%)</td>
<td>35.9 (47.1%)</td>
<td>23.7 (65.1%)</td>
<td>10.1 (85.1%)</td>
</tr>
<tr>
<td>Contribution to $\Delta A$ from $\Delta LWP$</td>
<td>-</td>
<td>3.8%</td>
<td>14.7%</td>
<td>39.9%</td>
<td>67.9%</td>
</tr>
<tr>
<td>A (%) using mean cloud cover and in-cloud LWP</td>
<td>70.0</td>
<td>52.8 (24.6%)</td>
<td>39.5 (43.6%)</td>
<td>27.2 (61.1%)</td>
<td>12.0 (82.9%)</td>
</tr>
</tbody>
</table>

*The number in the parentheses shows the relative difference from NC200. Contribution from the changes in liquid water path ($\Delta LWP$) to the relative difference in pseudo albedo ($\Delta A$) is indicated in the fourth row. Pseudo albedo computed using the mean cloud cover and in-cloud LWP is shown in the last row.
maintain a well-mixed circulation and surface moisture supply in the MBL. The deepening MBL and the large-scale subsidence also facilitate the entrainment of drier air from the free troposphere, stimulating evaporation of cloud water, and thus boundary layer stratification. Despite dramatic reduction, the amount of cloudiness and in-cloud LWP of nondrizzling clouds is still larger than that of drizzling clouds (Figures 1b and 1d). Therefore, our simulations demonstrate that low N_c is associated with a high precipitation rate, low cloud water, and low fractional cloudiness.

Our simulations confirm that drizzling stratocumulus correlate with small fractional cumulus clouds with lower cloud bases (Figure 2a), distinct from that of nondrizzling clouds. The cumulus-like clouds disappear in NC10-NE, implying the key role of subcloud evaporative cooling and moistening in reducing cloud bases, as well as building conditional instability, which allows cumulus to form on top of the surface mixing layer [Stevens et al., 1998]. Drizzle also leads to substantially shallower MBL (Figure 1c). The tendency of drizzle to suppress the MBL correlates with reduced cloud water and increased cloud droplet sedimentation which in turn results into reduced cloud top radiative [Ackerman et al., 1993] and evaporative cooling [Bretherton et al., 2007], and weaker turbulence (Figure 2b). Drizzle also stabilizes the boundary layer by giving up evaporative cooling of cloudy downdrafts to raindrop evaporative cooling below cloud bases [Stevens et al., 1998]. As a result, the MBL depth for NCS at the end of the simulation is about 200 m shallower than that of NC25 (Figure 1c), equivalent to a reduction of ~3 cm s^{-1} in entrainment rate, consistent with Ackerman et al. [2004].

As seen in Figure 1, the albedo correlates stronger with the in-cloud LWP than with the cloud cover, as a result of a much stronger reduction in the in-cloud LWP (at least three times stronger) than in the cloud cover with the decrease of N_c (Table 1). For example, the in-cloud LWP reduces by about a factor of 2 when N_c reduces from 200 mg m^{-1} to 10 mg m^{-1}, while the cloud cover decreases by 15% only. Note that despite large changes in albedo, no marked difference can be seen in terms of the cloud cover and the in-cloud LWP between NC200 and NC25 (Figures 1b and 1d) since their albedo difference is mostly attributed to the Twomey effect (Table 1).

### 4. Length Scale of Stratocumulus Organization

The cloud size distribution for different N_c can be examined qualitatively from the snapshots of albedo (Figure 3). Each of these stratocumulus cloud fields tends to have developed some organization, irrespective of the amount of drizzle, in agreement with previous studies (SS08). While the shape of the cells is slightly irregular in the albedo snapshots due to the presence of wind shear [Wang and Feingold, 2009], a more open cell-like structure is found with the increase of drizzle, which tends to be organized on a larger scale at 5 h (Figure 3).

Drizzling stratocumulus clouds bear different topology and dynamics from nondrizzling clouds (Figure 4). A comparison of NC10 and NC200 reflects a well-organized open cell structure in the presence of drizzle (surface precipitation rate ~ 0.7 mm d^{-1}), which features broad areas of weak vertical velocities in the center enclosed by narrow ring-shaped bands of strong positive upward motion. Although the cellular pattern of
anomalies can be identified in both NC200 and NC10, it is more intense and regular in the presence of drizzle. The heterogeneous spatial distribution of $\theta_l$ anomalies is also intensified with the development of drizzle. Comparing NC10 and NC10-NE suggests that the intensified $\theta_l$ variability in the subcloud layer correlates with drizzle evaporation in the lowest 300 m, without which the open cell network of $w$ and $q_t$ is less pronounced. This result suggests that the subcloud evaporation of drizzle makes clear changes to the subcloud topology, which is in general agreement with SS08. In the remainder of the section, the extent to which the subcloud thermodynamic structure regulates the cloud organization is explored.

To quantify the results from Figures 3 and 4, a two-dimensional (horizontal) spatial Fourier transform is applied to $q_t$ as a function of total wave number $k$, where $k^2 = k_x^2 + k_y^2$ and $k_x$ and $k_y$ are the wave numbers along the axis of the domain. The details of the methodology can be found in dR04. Figure 5 shows the variance spectra of $q_t$ in the subcloud layer (300 m) and cloud layer (700 m). The choice of $q_t$ is based on the analysis of dR04 who found that for stratocumulus, $q_t$ and $\theta_l$ are dominated by mesoscale fluctuations, with scales much larger than the order of boundary layer depth. The cloud patterns are mostly represented by the variance of $q_t$, thus, the variance spectra of $q_t$ are examined here. The corresponding spectral length scale $\Lambda_r = 1/k_r$ is calculated from each spectrum following dR04, in which $k_r$ is the critical wave number where two thirds of the variance resides at wave numbers larger than $k_r$. $\Lambda_r$ is analogous to finding the spectral peak with the advantage of being less sensitive to noise and thus is used here to indicate organizational scale.

Figures 5a and 5b show the variance spectra of $q_t$ in the subcloud and cloud layers for different $N_c$. The variance spectra of $q_t$ in the cloud layer agree qualitatively with those in the subcloud layer, which is in general larger for low $N_c$ cases due to the large dry-moist contrast in the presence of drizzle. In accordance with the visual inspection of albedo snapshots in Figure 3, both drizzling and nondrizzling stratocumulus show dominant peaks on mesoscale, with the corresponding spectral length scale of $q_t$ variance (Figures 5c and 5d) representing well the typical diameter of coherent structures occurring in the $q_t$.
anomaly field (e.g., Figure 5). For drizzling stratocumulus, lower $N_c$ correlates with a larger spectral length scale in the cloud layer for the first 10 h (Figure 5d). The discrepancy in scale between cases is less evident after 10 h since the removal of cloud water via drizzle accelerates the depletion of cloud water such that the precipitation rates for different $N_c$ cases converge after 10 h (Figure 1a). While drizzle tends to regulate the mesoscale development, it is not the only mechanism of mesoscale organization, as implied from the comparable cloud scale of nondrizzling stratocumulus in Figure 5d.

Compared with drizzle rates, the impact of MBL depth on cloud scale is less obvious. As seen in Figure 1c, the MBL depth of our simulations spans the range from ~700 m to ~1 km due to the tendency of precipitation to suppress the MBL growth. However, such diversity is not reflected in horizontal spectral length scales (Figure 5d). For drizzling stratocumulus, the horizontal scale tends to depend more on the precipitation rate which increases cloud scale even when the MBL depth is lower. For nondrizzling stratocumulus (NC200), cloud scale greatly increases after 8 h when the boundary layer is too deep to maintain well mixed. The scale of NC200 exceeds the domain size after 12 h. While it is interesting to explore the impact of a stratified MBL on cloud scales, it is beyond the scope of this study and thus we defer it for future research.

Figure 4. Cross sections of instantaneous perturbations from the horizontal mean values of $w$, $\theta_l$ (with contours of $q_r = 0.01$ g kg$^{-1}$ superimposed), and $q_r$ fields at the 100 m level at 5 h for NC200, NC10, and NC10-NE.
When raindrop evaporation below 300 m is omitted (NC10-NE), the spectral density of subcloud $q_t$ decreases at all scales but faster at scales larger than 1 km (Figure 5a). The weaker $q_t$ variance at large scales underlines the variance at smaller scales and thereby decreases the length scale calculated based on spectrum integral (Figure 5c). The less coherent subcloud temperature and moisture are visually confirmed from the $\theta_l$ and $q_t$ anomaly field shown in Figure 4. In the cloud layer, the change of $q_t$ variance for NC10-NE is insignificant (Figures 5b and 5d). The comparable large-scale $q_t$ variances in the cloud layer for NC10 and NC10-NE suggest that subcloud moisture and temperature variances do not contribute to those in the cloud layer.

To evaluate the impact of the subcloud scale, we performed filtered runs (see section 2) of NC200 and NC10 (Figures 6 and 7). Comparing Figures 4 and 6, one can observe that the filtering procedure worked properly to remove the large-scale fluctuations from the subcloud temperature and moisture fields. This is more quantitatively seen from the variance spectra shown in Figure 7. The subcloud $q_t$ variance at scales larger than 0.2 km decreases noticeably for both filtered runs (Figure 7a), reducing their subcloud length scales by a factor of 2 (Figure 7c). Note that the amplification of the variance at scales smaller than 0.2 km (Figure 7a) is due to the inhibition of turbulence cascade to large scales.

For NC200, the removal of large scales in the subcloud layer greatly reduces its length scale in the cloud layer (Figures 7b and 7d). The reduction of the in-cloud length scales is visualized by comparing the albedo snapshots in Figures 3a and 6a. By contrast, the absence of large scales in the subcloud layer of NC10 barely affects the cloud field, as seen in Figures 6b, 7b, and 7d. This confirms that mesoscale organization of drizzling stratocumulus is independent of subcloud moisture and temperature variance. Our results also imply that the spatially inhomogeneous surface latent heat and sensible heat fluxes are not essential for the mesoscale
organization of drizzling stratocumulus. This is in agreement with Kazil et al. [2014] who found that the spatial distribution of surface fluxes has little bearing on cloud properties in stratocumulus organization.

5. Decrease of the Amount of Cloudiness for Drizzling Clouds

Looking back at Figures 1a and 1b, we observe that low $N_c$ modifies the cloudiness through drizzle. This is reflected when the surface and cloud base precipitation rates are plotted against the cloud cover (Figure 8a).
A threshold of surface precipitation rate around 0.6 mm d\(^{-1}\) is observed, below which the amount of cloudiness decreases dramatically fast. The behavior of the cloud cover suggests that there might exist an \(N_c\) (or precipitation) threshold above which the cloud cover remains high and is insensitive to the aerosol loading, maintaining closed cells, while open cells are very likely to occur with only a slight reduction of the \(N_c\) (or precipitation). The threshold behavior is also observed in previous modeling studies [e.g., Ackerman et al., 2003; Berner et al., 2013] and is supported by observations of VOCALS Regional Experiment [e.g., Wood et al., 2011], which shows that the precipitation rates in the closed cell regions are not much lower than that in POCs.

To understand the role of precipitation in reducing cloud cover, we quantify in each model column at 5 h, the rate of change in LWP (% min\(^{-1}\)) caused by cloud droplet condensation and cloud droplet evaporation (Figure 8b) and by autoconversion and accretion (Figure 8c). It is shown that the cloud droplet evaporation overwhelms condensation and dominates in the low-LWP columns, while the formation of raindrops is more efficient in the high-LWP columns. This suggests that cloud droplet evaporation is responsible for depleting the thinnest clouds and thus the reduction of cloud cover, while removal of cloud water by drizzle does not directly contribute to the decrease of cloudiness.

However, with sufficient drizzle, the cloud water is efficiently depleted and the cloud thickness of high-LWP columns reduced [Ovchinnikov et al., 2013], contributing to an inhomogeneous spatial distribution of LWP as revealed in Figure 8d. The highest frequency (or count) of LWP shifts toward the low-LWP bins with the increase of precipitation. Note that the discontinuous high frequency at the extremely low LWP (0–10 g m\(^{-2}\)) in the presence of drizzle [Yamaguchi and Feingold, 2015] is attributed to the newly formed isolated low clouds with cloud top height on average lower than 750 m and is therefore not the focus of this study. For moderately drizzling cases (i.e., NC15 and NC25), the broad peak at large LWP suggests that the

Figure 7. Same as Figure 5 but for filtered and unfiltered runs of NC10 and NC200.
The major portion of the clouds maintains thick and is less affected by the cloud droplet evaporation. When surface precipitation rate is larger than 0.6 mm d\(^{-1}\) (or \(N_c\) less than 15 mg\(^{-1}\) in our simulations), the broad peak of LWP transitions to low-LWP bins, leaving the largest area of the thinnest clouds to the evaporation depletion.

Although the column-based cloud droplet evaporation is less efficient for drizzling clouds (Figure 8b) due to the weaker entrainment as implied from the shallower boundary in Figure 1c, the large area of the thinnest clouds promotes the efficient depletion of the cloud water and therefore an abrupt drop in the cloud cover (Figure 8a). This suggests the critical role of heterogeneous LWP distribution in increasing cloud droplet evaporation and in reducing the amount of cloudiness, which supports \textit{Wood and Hartmann} [2006] who found that the cloud cover is strongly linked with the LWP variability over the eastern subtropical oceans.

### 6. Mean Cloud Properties on Insufficient Scales

In the previous sections, evidence was presented demonstrating the influence of drizzle to the spectral length scale in the cloud field. One may therefore ask how important the emergence of large-scale variance is to the mean properties. By reducing the domain size (with the same horizontal grid resolution as for the large domain simulation), it is shown in Figure 9 that an insufficient scale biases the clouds toward lower
amounts of cloudiness and liquid water at the end of the simulations. This is likely due to the forced circulation that generates additional downdrafts, resulting in excessive cloud top entrainment and substantial warming of the MBL. The MBL warming facilitates evaporation and leads to a drop in liquid water. In general, NC200 shows similar results to NC10 except that the MBL depth does not increase much in small domains (Figure 9). This is attributable to the sharp decrease in cloud water via the excessive entrainment, which substantially reduces cloud top cooling and in turn weakens the entrainment, compensating for the gain.

7. Discussion and Conclusions

Large-eddy nocturnal simulations of stratocumulus with a range of fixed cloud droplet number concentrations from 5 mg$^{-1}$ to 200 mg$^{-1}$ are presented. The simulations are performed on a large domain (25.6 km × 25.6 km) to allow mesoscale development of clouds.
When horizontal length scale is not sufficient for mesoscale organization, the mean properties of the cloud field are found biased toward less amount of cloudiness as a result of additional entrainment via forced circulation.

We have shown that drizzle plays an important role in changing the amount and the spatial variability of cloud water, as well as pseudo albedo. We found that between stratocumulus in which little to no precipitation reaches the surface, the Twomey effect explains most of their pseudo albedo difference. However, for stratocumulus with sufficient drizzle, the hydrological cycle effect plays a more significant role. The in-cloud LWP in the presence of drizzle decreases 3 times faster than the amount of cloudiness, dominating the changes of pseudo albedo. Since our study focused on nocturnal stratocumulus clouds, we cannot use it for a quantitative assessment of the Twomey effect. However, our result remains highly suggestive and provides incontrovertible evidence that drizzle plays an essential role in determining cloud radiation.

The simulations indicate that while drizzle evaporation below 300 m changes the topology and dynamics in the subcloud layer, its contribution to the moisture variance in the cloud layer is not evident. This is confirmed by a filtering analysis that removed the large scales of thermodynamic quantities in the subcloud layer. The filtering procedure failed to reduce the horizontal length scale for drizzling stratocumulus, suggesting that the mesoscale organization of drizzling stratocumulus is independent of the subcloud moisture and temperature variance. This further implies that the spatial distribution of surface fluxes is not essential for cloud mesoscale variability. Although drizzle evaporation below 300 m does not change the mean properties and the mesoscale organization of the cloud layer, it tends to extend the lifetime of stratocumulus by a couple of hours due to the cooling of the subcloud layer. This, however, is not consistent with Albrecht [1989] who argued that drizzling stratocumulus clouds tend to have a shorter lifetime and more easily break up than nondrizzling stratocumulus due to the stabilized MBL induced by drizzle.

We stress that the focus of this study is marine stratocumulus associated with a strong inversion layer. It is important to recognize that the organized open cellular stratocumulus clouds are different from the trade wind cumulus [Xue et al., 2008; Seifert and Heus, 2013] and from the deep convective clouds [e.g., Boëng et al., 2012]. As is discussed in Seifert and Heus [2013], the cold pools in trade wind cumulus and deep convective clouds often have pronounced dry cores, while this is not the case for open cellular stratocumulus [Zuidema et al., 2012; SS08]. Therefore, the results shown in this study do not represent the trade cumulus and deep convective clouds. The difference between these cloud regimes warrants further investigation.

Our simulations suggest that while drizzle does increase cloud scale, it is not the only mechanism of mesoscale organization. The nondrizzling clouds exhibit a comparable scale to the drizzling clouds during the first 8–10 h of the simulation. Their length scales increase significantly afterward when the MBL is too deep to maintain well mixed. The fast growth of nondrizzling cloud scale suggests that a stratified MBL tends to broaden cloud horizontal scales. This result also supports Wood and Hartmann [2006] who provided observational evidence of a remarkable correlation between MBL depth and horizontal scale of the mesoscale cellular convection. For drizzling stratocumulus, the correlation between MBL depth and cloud horizontal scale is not evident in our study. The horizontal scales of drizzling stratocumulus appear to depend more on precipitation and converge in later hours of the simulations despite the diverse MBL depths. One possibility is that a greater discrepancy in MBL depth or longer time is required for the impact of MBL depth to be seen. Another interpretation is that the MBL depth is less likely the major control on horizontal cloud scale but more likely the mark of the position of clouds. Since the MBL over the eastern subtropical ocean deepens [Zhou et al., 2015] with the increase of the sea surface temperature [Wyant et al., 1997; Sandu and Stevens, 2011], a deeper MBL indicates a position farther away from the coast. In a Lagrangian perspective, this allows stratocumulus to evolve for longer along the easterly trade winds. It is likely that the cloud-scale development might merely respond to the evolution of the flow, which is independent of MBL depths. Furthermore, the interaction between closed cells and POCs might alter the MBL depth and their length scale behaviors. A recent observation shows that difference of MBL depths between the closed cells and POCs is usually not significant [Wood et al., 2011], presumably due to the secondary circulations near the cloud top that diverts subsiding free-tropospheric air away from the POC into the surrounding closed cell region [Bretherton et al., 2010; Berner et al., 2011]. To better understand the influence of MBL depths, future studies are required.

An $N_c$ (or precipitation rate) threshold is found in our study below which the amount of cloudiness decreases dramatically fast. By performing water budget analyses, we showed that the profound
depletion of the cloudiness is due to the spatial variability of clouds induced by drizzle. The large portion of the thinnest clouds promotes evaporation depletion of cloud water despite weaker entrainment. This result explains the findings of Wood and Hartmann [2006] that cloud cover is strongly linked with the LWP variability over the eastern subtropical oceans. It is noteworthy that such threshold of \( N_c \) (or precipitation rate) is likely affected by meteorological variability. For instance, a drier free troposphere, a deeper boundary layer, or a weaker inversion likely makes the threshold effect sharper via the increase of cloud droplet evaporation. A prognostic \( N_c \) also tends to sharpen the effect due to the precipitation scavenging of CCN.

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