Mechanisms of Convective Clustering During a 2-Day Rain Event in AMIE/DYNAMO

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Abstract Physical mechanisms that are key to observed convective clustering in 2-day rain events are examined. Previous analysis of the 2-day rain events during the Atmospheric Radiation Measurement Madden-Julian Oscillation Investigation Experiment (AMIE)/Dynamics of the Madden-Julian Oscillation (DYNAMO) field campaign data revealed two distinct phases of convective clustering. Using a cloud-system-resolving model, we perform a series of intervention experiments to investigate the underlying mechanisms for convective clustering in each phase. In the developing phase, in addition to previously emphasized processes such as the cold pool–updraft interaction and moisture-convection feedbacks, our results show that the vertical wind shear in the lower free troposphere is a critical factor for convective clustering. Stronger lower free-tropospheric wind shear increases the entrainment of environmental air into updrafts and prevents convective clouds from being omnipresent. This result suggests that stronger vertical wind shear in the lower free troposphere can help spatially organize the convection, even for non-squall-line-type convective systems. In the decaying phase, the cold pool–updraft interaction becomes less effective in aggregating convective clouds because the boundary layer is widely cooled by stratiform precipitation. Instead, the mesoscale downdraft driven by the stratiform precipitation becomes the dominant factor to maintain the relatively aggregated convection. Additionally, removing horizontal variations in radiative heating has no impact on convective clustering on this 2-day time scale, even in the decaying phase when stratiform clouds are widespread. The implication of these results for improving the representation of mesoscale convective organization in convection schemes is discussed.

Plain Language Summary Tropical thunderstorms often cluster together. Studies have suggested that the degree to which the tropical thunderstorms are clustered impacts Earth’s energy balance and water cycle, as well as extreme precipitation events. However, the processes controlling the spatial distribution of the thunderstorms are poorly understood and are not properly represented in most computer models for weather and climate prediction. This study aims to understand the processes that control the organization (i.e., clustering) of thunderstorms. We target an observed 2-day rain event over the tropical ocean and investigate its organization processes by conducting a series of numerical experiments that are designed to test selected physical mechanisms. We show that the key organization processes may differ depending on the life stage of the rain event. During the intensifying stage where thunderstorms are continuously forming, the moisture distribution and the environmental vertical wind shear act to confine the thunderstorms within the moist regions. Changes in boundary layer temperature driven by the thunderstorm further trigger additional thunderstorms nearby, leading to more organized thunderstorms. During the decaying stage where thunderstorms are gradually dissipating, the remaining clustered thunderstorms will drive circulations that help maintain the clustered thunderstorms.

1. Introduction

Organized tropical convective systems significantly impact Earth’s climate system. For example, more spatially organized convection is suggested to be associated with a drier troposphere and stronger outgoing long-wave radiation with the same amount of precipitation (Tobin et al., 2012, 2013). In other words, more organized convection corresponds to a higher convective precipitation efficiency, which controls a large sensitivity of top-of-atmosphere net cloud radiative forcing that is suggested to be an important factor for climate sensitivity (Lutsko & Cronin, 2018; Zhao, 2014). Accurately representing the spatial organization of convection in global climate models (GCMs) is therefore necessary for them to accurately predict the future state of Earth’s climate system.
The most common type of organized tropical convective systems on the scale of approximately 100 km is the mesoscale convective system (MCS; Houze, 2004), which most contemporary GCMs cannot resolve with their native grid boxes. Many studies have emphasized the need for parameterizing the effects of MCSs in convection schemes (e.g., Schumacher and Houze, 2004; Choudhury & Krishnan, 2011; Feng et al., 2018). In particular, a few recent studies have demonstrated that explicit representation of mesoscale convective organization is necessary for a GCM to realistically simulate both the mean state and tropical variability (e.g., Ahn et al., 2019; Mapes & Neale, 2011). Unfortunately, however, representation of the effects of MCSs is missing in most convection schemes (Del Genio et al., 2015). This lack of organized convection in convection schemes is partly due to the lack of complete understanding of the mechanisms leading to spatially organized convection under various environmental conditions (Mapes & Neale, 2011).

A recent observational study of tropical oceanic convection provided new insights into the mechanisms of convective clustering on a time scale of a few days by investigating the spatial patterns of convective clouds and their evolution during a series of 2-day rain events. Cheng et al. (2018; hereafter CKR18) quantified the spatial evolution of convective echoes in the S-PolKa radar data set collected during the Atmospheric Radiation Measurement Madden-Julian Oscillation Investigation Experiment (AMIE)/Dynamics of the Madden-Julian Oscillation (DYNAMO) field campaign. CKR18 used \( I_{\text{org}} \), a scalar metric previously used for cloud-resolving model-simulated deep convection (Tompson & Semie, 2017), to quantify the spatial clustering of convection in the ground-based S-PolKa radar data set. They found that convection becomes more spatially clustered during 2-day rain episodes from 10 hr before to 10 hr after the time of peak rain rate. This result is consistent with that of Zuluaga and Houze (2013) who showed that the dominant cloud type during the same 2-day rain episodes evolves from shallow convective to deep convective, deep and wide convective, and then to stratiform clouds.

Through analyzing the evolution of \( I_{\text{org}} \) and the number of convective clouds (\( N \)) during the 2-day rain events, CKR18 separated the entire period of convective clustering (i.e., increasing \( I_{\text{org}} \)) into two distinct phases. The first phase approximately corresponds to the time period between 10 hr before to the time of peak rain rate, during which the number of convective clouds is increasing. During this developing stage of the 2-day rain events, newly formed convective clouds appear preferentially near existing ones, leading to an increase in \( I_{\text{org}} \). This period, which is also characterized by increasing \( N \), is referred to as Phase 1 by CKR18. The second phase approximately corresponds to the time period between the time of peak rain rate to 10 hr after, during which the number of convective clouds (\( N \)) is decreasing. During this phase, the relatively aggregated groups of convective clouds that are often embedded in stratiform clouds are sustained longer than the typical lifetime of individual cumuli. As a result, the convective system consists of a few aggregated convective clouds and reaches a higher degree of convective clustering by the end of this decaying stage of the 2-day rain events. This later period, which is characterized by an increasing trend of \( I_{\text{org}} \) and a decreasing trend of \( N \), is referred to as Phase 2 in CKR18.

While CKR18 revealed two distinct phases of convective clustering during the 2-day rain events, the mechanisms through which convection becomes clustered in each phase remain unclear. CKR18 proposed possible mechanisms for convective clustering based on the observed spatial evolution of convective systems. They hypothesized that the updraft-cold pool interactions were one of the key mechanisms of convective clustering in both phases. That is, the new convection can be triggered near the existing convective clouds via the additional lifting along the cold pool edges in Phase 1, and the cold pools under stratiform clouds may restrict the area available for convection, helping convection become more localized in Phase 2. CKR18 also hypothesized that the stratiform-driven mesoscale downdraft was an important factor for convective clustering during the Phase 2 period.

In this study, we test the hypothesized mechanisms for convective clustering during the AMIE/DYNAMO 2-day rain events using cloud-system-resolving model simulations of one representative event. In particular, we will perform a series of intervention experiments to address the following questions: (1) What are the dominant mechanisms of convective clustering during the observed 2-day rain events? (2) Are the dominant mechanisms the same or different between Phase 1 and Phase 2?

In addition to testing mechanisms related to cold pools and mesoscale downdrafts, we will also explore the relevance of convective self-aggregation in radiative convective equilibrium (RCE) in the 2-day rain events. Held et al. (1993) discovered in a 2-D RCE simulation that random convective clouds tend to cluster over...
time and become aggregated into a single cluster, which was confirmed in later studies (e.g., Bretherton et al., 2005; Tompkins, 2001). More recent studies have suggested the important role of radiation in this self-aggregation (e.g., Muller & Held, 2012; Wing & Emanuel, 2014; Muller & Bony, 2015; Coppin & Bony, 2015; Arnold & Putman, 2018), while others point to moisture-convection feedbacks (Tompkins, 2001; Mapes & Neale, 2011; Colin et al., 2019) and generation of available potential energy (Yang, 2018, 2019) as being the dominant mechanism. However, it is not yet clear how convective self-aggregation found in idealized simulations can be related to organized convection in the real world (Wing et al., 2017). The typical time scale of convective self-aggregation is longer than 10–30 days. Whether the mechanisms to explain convective self-aggregation in RCE can be applied to mesoscale convective clustering that has shorter temporal scales as in the 2-day rain events remains relatively unexplored and will be included as part of our intervention experiments.

The next section describes the observational data set, model configuration, and the method we use to quantify convective clustering. The design of each intervention experiment is described in section 3. The modeling experiment results are presented in section 4. Section 5 discusses the relative importance of each mechanism for Phases 1 and 2, and conclusions and broader implications of this work are described in section 6.

2. Data, Model, and Method

2.1. Data

The National Center for Atmospheric Research's Doppler, dual-polarization S-PolKa radar was deployed on the Addu Atoll (0.6°S, 73.1°E) in the Maldives during the AMIE/DYNAMO field campaign. The reflectivity field used in this study is from S-PolKa’s S-band 360° surveillance scans, collected at eight elevation angles ranging from 0.5° to 11°, which were then interpolated to Cartesian coordinates with 0.5° horizontal resolutions. Radar reflectivity at 2.5-km height is used in this study. The radar data were collected every 15 min from 1 October 2011 through 15 January 2012. During this period, ten 2-day rain events were observed (see table 2 in CKR18). The characteristics of convective clouds during these events are documented in previous literature, in terms of the frequency of different cloud types (Zuluaga & Houze, 2013), diurnal cycle (Rowe et al., 2019), and the evolution of spatial clustering of convection (CKR18).

2.2. Weather Research and Forecasting Model Configurations and the Target Case

The model used for the simulations is the Weather Research and Forecasting (WRF) model version 3.8.1 (Skamarock et al., 2008). The horizontal size of the model domain is 256 km × 256 km with doubly periodic horizontal boundary conditions. The horizontal grid spacing is 1 km, and there are 62 levels in the vertical with the model top at z = 25 km, with stretched spacing in the range of 40 to 2,500 m from the lowest to the highest levels. The model time step is 10 s. The Yonsei University boundary layer scheme (Hong et al., 2006; Noh et al., 2003), Rapid Radiative Transfer Model for GCMs radiation physics (Iacono et al., 2008), and Thompson microphysics (Thompson et al., 2008) are used with no cumulus parameterization.

In this paper, we focus on the AMIE/DYNAMO 2-day rain event starting from 0800 UTC 15 October 2011 through 0800 UTC 17 October 2011. This event is selected because (1) the time evolution of the clustering metrics during this event most closely resembles the composite of all 2-day rain events (CKR18) and (2) the convection was mostly locally generated—there was no large preexisting convective system that passed through the radar domain. The simulations for this event using the default configuration will be referred to as Control hereafter, in which the evolution of the observed convective clustering is realistically captured (see section 4.1). To ensure the robustness of the results, we run five-member ensemble simulations for all of the experiments used in this study. Five ensemble members are initialized at 1800 UTC and 2100 UTC on 13 October and 0000 UTC, 0300 UTC, and 0600 UTC on 14 October, which allows a spin-up period of at least a day before the targeted 2-day rain event.

All ensemble members were forced by the same large-scale forcing data set, created using the sounding data at Gan Island combined with the ECMWF analyses (Zhang et al., 2001; Zhang & Lin, 1997). Horizontal wind fields are relaxed toward the forcing wind fields for all model grids with a relaxation time scale of 1 hr. For temperature and moisture, the large-scale horizontal advective tendencies are specified from the large-scale forcing data set. The vertical advection terms are explicitly evaluated by the WRF advection scheme, with
vertical motion profiles being specified by those in the large-scale forcing data set. This approach differs from those used in similar studies (e.g., Li et al., 2018; Wang et al., 2015) in that in the latter cases the horizontal average of temperature and moisture in the domain is used for large-scale advection. A sensitivity test using these two different settings reveals that the way of imposing large-scale vertical velocity does not affect our main results (not shown).

Figure 1 shows the large-scale forcing fields used for the simulations. The 0 hr is defined as 0800 UTC 16 October, which is the time of the maximum radar-domain mean rain rate after applying a 24-hr running average. Figures 1a–1c show the three-dimensional wind field from the large-scale forcing data set. The horizontal wind component is mostly in the zonal direction, with a rather weak meridional wind component throughout the event. From −24 to +12 hr, there is easterly vertical wind shear of about 10 m/s between $z = 1$ and 5 km. From −6 to 0 hr, there is large-scale ascent in the lower free troposphere, which would favor convective development. The large-scale forcing also gradually destabilizes the environment from about −16 to −3 hr, as indicated by the positive potential temperature horizontal advection near the surface and the positive horizontal moisture advection in the lower free troposphere (Figures 1d and 1e). After the time of the peak rain rate, the environment becomes gradually stabilized, as indicated by the negative potential temperature tendency near the surface from 0 to +6 hr and the negative moisture tendency in the lower free troposphere from −6 to +15 hr (Figures 1d and 1e).

2.3. Convective Clustering Quantification

As in CKR18, we use the Organization Index $I_{org}$ (Tompkins & Semie, 2017) as a scalar metric to represent the degree of convective clustering in the cloud-system-resolving model simulations. $I_{org}$ is calculated based on the Nearest-Neighbor Cumulative Distribution Function (NNCDF; Weger et al., 1992; Zhu et al., 1992) of convective objects. To obtain an $I_{org}$ value from a 2-D radar reflectivity field, we first identify the convective objects using a convective-stratiform classification algorithm (detailed below). The nearest-neighbor distance is then calculated for every convective object to obtain a cumulative distribution function.
Table 1  
List of Experiments

| Experiment name | Forcing | Forced level | Intention |
|-----------------|---------|--------------|-----------|
| BLT             | \( \delta_t \overline{T}(x,y,z,t) = \left( T(x,y,z,t) - \overline{T}(z,t) \right) / \tau \) | All grid points within boundary layer (lowest model layer to 0.8 km) | To weaken the interactions between boundary layer cold pools and updrafts |
| LFTqv           | \( \delta_t \overline{Q_v}(x,y,z,t) = \left( Q_v(x,y,z,t) - \overline{Q_v}(z,t) \right) / \tau \) | All non-cloudy grid points in lower free troposphere (0.8 to 5 km) | To weaken the moisture-convection feedback and mesoscale circulation |
| BLWS            | \( \delta_t \overline{U}(x,y,z,t) = \left( U(x,y,z,t) - \overline{U}(z,t) \right) / \tau \) | All grid points in boundary layer (lowest model layer to ~0.8 km) | To weaken the boundary layer wind shear |
| LFTWS           | \( \delta_t \overline{V}(x,y,z,t) = \left( V(x,y,z,t) - \overline{V}(z,t) \right) / \tau \) | All grid points in lower free troposphere (0.8 to 5 km) | To weaken the lower free-tropospheric wind shear |
| RAD             | \( \delta_t \overline{T}_{rad}(x,y,z,t) = \left( dT / dt \right)_{rad}(z,t) \) | All levels | To weaken the inhomogeneity of the radiative effect |

(i.e., NNCDF). The last step is to compare the obtained NNCDF to an idealized NNCDF that is calculated assuming all the convective objects are randomly spatially distributed over the same domain. The idealized NNCDF can be described by a Poisson CDF, given by a Weibull distribution (Stoyan et al., 1987; Weger et al., 1992):

\[
\text{NCDF}_{\text{random}}(r) = 1 - \exp(-\lambda \pi r^2),
\]  

where \( \lambda \) is the density of convective elements per unit area, (if \( N \) is the number of convective objects and \( L \) is the length of the domain, \( \lambda = N/L^2 \)), and \( r \) is nearest-neighbor distance.

In a numerical simulation, convective objects can be identified using a vertical velocity threshold as in Tompkins and Semie (2017). However, in order to have an apple-to-apple comparison with the observations, we choose to follow CKR18 and use the contiguous convective echoes (CCEs) as the convective objects. The convective echoes are identified by applying a rain-type classification algorithm to the simulated reflectivity field at the 2.5-km level. The reflectivity field in the WRF model is calculated using the mixing ratios of the hydrometeors in the simulations by the built-in subroutines. The rain-type classification algorithm of Powell et al. (2016) is used to classify each grid point into six categories: convective, isolated convective core, isolated convective fringe, stratiform, uncertain, and weak echoes. The parameters for the rain-type classification are based on the DYNAMO legacy data set (http://dynamo.fl-ext.ucar.edu/rsmas/dynamo_legacy/; see table 3 in CKR18). Note that the isolated convective core category is counted as convective to include the isolated/shallow convection in our analysis as in CKR18. Once the convective pixels are identified, they are grouped into CCEs by the condition that two pixels sharing a common side are identified as the same CCE. CKR18 showed that the number of CCEs can be used as an indicator of the overall activeness of convection. In the rest of the paper, as in CKR18, we will use the two scalar metrics—the number of CCEs (\( N \)) and \( L_{\text{avg}} \)—to characterize the spatial distribution of the convection.

3. Design of the Intervention Experiments

To examine the mechanisms responsible for convective clustering seen in the Control simulation and observations, a series of intervention experiments are designed. Based on existing knowledge of convective clustering, our intervention experiments target four fields: boundary layer temperature, lower free-tropospheric moisture, boundary layer and lower free-tropospheric vertical wind shear, and radiative heating. Experimental designs of each experiment are summarized in Table 1 and detailed in the rest of this section.

3.1. Boundary Layer Temperature (BLT)

Tropical oceanic convection may organize through the interaction of boundary layer cold pools and convective updrafts (e.g., Böing et al., 2012; Feng et al., 2015; Khairoutdinov & Randall, 2006; Rowe &
Houze, 2015; Zuidema et al., 2017). As raindrops fall through deep convective clouds, they evaporate and descend to the surface forming “cold pools” due to reduced buoyancy. The cold pools spread out horizontally near the surface, triggering new convective updrafts along its leading edges (Böing et al., 2012; Khairoutdinov & Randall, 2006).

To examine the possible role of cold pool-updraft interactions in the 2-day rain events, the boundary layer temperature perturbations are weakened by nudging the temperature in the boundary layer at every grid point toward its domain mean vertical profile at each time step using the following equation:

\[
\frac{\partial}{\partial t} \bar{\Theta}(x, y, z, t) = -\left( \frac{\Theta(x, y, z, t) - \bar{\Theta}(z, t)}{\tau} \right),
\]

where \(\bar{\Theta}\) denotes the horizontal domain average and \(\tau\) is the nudging time scale. \(\tau\) is set to 10 min in all simulations; varying \(\tau\) from 5 to 30 min does not affect our results (not shown). For this BLT experiment, \(\Theta\) in equation (2) represents the boundary temperature field, and the nudging term is applied only to the boundary layer, which is defined as the lowest 10 model layers, corresponding to the boundary layer top of approximately 0.8 km.

This experiment design is somewhat different from conventional approaches to weaken the boundary layer cold pools, in which the evaporative cooling is turned off (e.g., Jeevanjee & Romps, 2013; Muller & Bony, 2015; Yang, 2018). The intention of our approach is to focus on the lifting effect of the cold pools. While turning off evaporative cooling can achieve the same goal via essentially preventing the formation of cold pools, the domain mean temperature and moisture profile will be altered, which may lead to an undesired change to the overall convective characteristics in the domain (Böing et al., 2012; Kurowski et al., 2018).

It is noted that suppressing boundary layer temperature perturbations may also inevitably alter the formation of boundary layer circulation patterns via modifying the temperature and pressure gradients. As the boundary layer temperature fields are nudged toward the domain mean, the boundary layer circulation pattern would not be able to develop (see supporting information Figure S1). Yang (2018) suggested that the formation of the boundary layer circulation patterns is an essential process for convective clustering on the time scale >5 days. While it is not yet clear whether such boundary layer circulation patterns are critical for the convective clustering on the time scale of ~1 day, we note that that possibility cannot be excluded.

3.2. Lower Free-Tropospheric Moisture (LFTqv)

Numerous studies have shown that tropical oceanic convection is tightly coupled to column water vapor (Bretherton et al., 2004; Brown & Zhang, 1997; Holloway & Neelin, 2009; Neelin et al., 2009; Peters & Neelin, 2006; Rushley et al., 2018; Sherwood, 1999; Wang & Sobel, 2012). This relationship has been attributed to the lateral mixing around the convective updrafts; when convective updrafts develop within a relatively dry area, the entrainment of environmental dry air into the convective clouds effectively lowers the buoyancy, thereby weakening the convection. As a result, deep convective clouds form preferentially within the moist patches.

The experiment LFTqv is designed to test the possible role of fluctuations in the moisture field on convective clustering. Similar to BLT, equation (2) is added to the model moisture prognostic equations, where \(\Theta\) represents the lower free-tropospheric moisture field, to suppress lower free-tropospheric moisture fluctuations. The nudging term is applied to the lower free troposphere, which corresponds to the layer between the 11th to 30th model layer (approximately \(z = 0.8 \text{ km}\) to \(z = 5 \text{ km}\)). Only non-cloudy grid points, defined as the grids with cloud water vapor mixing ratio less than 0.1 g/kg everywhere in the column, are used to obtain the domain mean and are affected by the nudging term to avoid causing drastic changes in cloud properties.

3.3. Boundary Layer and Lower Free-Tropospheric Wind Shear (BLWS and LFTWS)

The environmental vertical wind shear has been suggested to aid convective clustering in certain circumstances. When the acceleration of boundary layer cold pools reaches the balance with the vertical wind shear, the updraft can be continuously generated near the original convective entity that produced the cold pools, thus increasing spatial clustering (Rotunno et al., 1988, hereafter RKW). Indeed, it has been found that in idealized numerical simulations, the vertical wind shear is critical for the development of MCSs, as weakening the vertical wind shear in simulations often prevents the formation of MCSs
(Anber et al., 2014; Fovell & Ogura, 1989; Weisman, 1992). However, these studies focus mostly on mid-latitude squall-line systems. The extent to which the RKW theory can be applied to tropical oceanic convection stay relatively unexplored. A set of experiments is therefore designed in this study to examine whether the vertical wind shear is critical to convective clustering during the 2-day rain events. To reduce the domain-averaged vertical wind shear, the following nudging terms are added to the momentum prognostic equations:

$$\partial_t u(x, y, z, t) = -\left(\overline{U(t)} - \overline{U(t)}\right),$$  

(3)

where $u$ represents the zonal and meridional wind fields and $\overline{()}$ denotes the vertical mass-weighted average. The role of these nudging terms is to relax the domain mean vertical wind shear toward the vertical mass-weighted domain mean wind ($\overline{U(t)}$).

It is important to note that the vertical wind shear between different height levels may have different influences on the organization of convection. In a study of a continental MCS, Chen et al. (2015) showed that increasing wind shear in the low-to-middle troposphere (0-5 km) leads to a more organized quasi-line convective system, while increasing wind shear at the upper levels (>10 km) has minimal effect on the organization of the convection. Anber et al. (2014) showed that tropical convection is also sensitive to the depth of vertical wind shear. They found that the midlevel shear is optimal for producing maximum mean precipitation.

To differentiate the effect of different levels’ environmental wind shear on convective clustering, two experiments are conducted in this study, where the nudging term in equation (3) is applied only to the boundary layer (BLWS) and only to the lower free troposphere (LFTWS). The definitions of the boundary layer and lower free troposphere follow those in BLT and LFTqv.

It is worthwhile to note that relaxing the model wind field toward the mass-weighted domain mean wind can also change the advection of moisture, which might affect the evolution of clouds in the simulation. However, we found that the moisture advection that would be modified in LFTWS (i.e., those caused by the domain mean wind) takes only a small portion (<10%) of the total moisture advection in the Control simulation (not shown), suggesting that the changes in the moisture advection might not be responsible for the difference between Control and LFTWS.

3.4. Radiative Heating (RAD)

Recent studies have shown that in idealized RCE simulations, convective clouds tend to form within a localized wet region while the rest of the domain is covered by a dry patch, which is referred to as convective self-aggregation (e.g., Bretherton et al., 2005; Held et al., 1993; Wing et al., 2017). Anber et al. (2016) also demonstrated that radiative heating may complicate the effects of vertical wind shear on mesoscale organization and precipitation.

In our RAD experiment, the radiative heating of every grid points is made equal to the domain-averaged radiative heating at each level, which can be equated as

$$\overline{(\partial_T \rho)_{\text{rad}}(x, y, z, t)} = \overline{(\partial_T \rho)_{\text{rad}}(z, t)},$$  

(4)

where $(\partial_T \rho)_{\text{rad}}$ denotes the sum of shortwave and longwave radiative heating rates. By applying equation (4) to every grid point in the domain, the radiative heating becomes horizontally uniform at every time step. This prevents any horizontal gradient in radiative heating and the associated radiative-driven circulation, which was suggested as an important driver for convective self-aggregation in the idealized RCE simulations (Bretherton et al., 2005; Muller & Bony, 2015; Muller & Held, 2012; Wing & Emanuel, 2014).

However, it is important to note that the typical time scale of the convective self-aggregation in the RCE simulations (O(10 days)) is much longer than the time scale of the convective clustering during the 2-day rain events (~1 day). Additionally, our simulations are forced by the observed large-scale forcing, which is fundamentally different from the self-aggregation studies. Hence, it should not be taken for granted that the mechanisms found for self-aggregation are functioning in the simulations of this study. Nevertheless,
it is interesting to examine whether the convective self-aggregation mechanisms found in an idealized framework are also functioning in a more realistic rain event simulation.

4. Experiment Results

4.1. Control Simulation

Figures 2a to 2d show the observed evolution of the 2.5-km S-PolKa radar reflectivity field during the 16 October case. Note that the time information in this figure refers to the hour relative to the time of the peak rainfall. The spatial distribution of convection (i.e., CCEs) evolves as follows. At −20 hr, the domain is mostly covered by clear sky except for a few small CCEs that are aligned along a linear system (not shown). From −20 to −6 hr, as the large-scale environment becomes moister and less stable (Zuluaga & Houze, 2013, CKR18), more convective clouds form over a wider region of the domain. During this period, \( I_{\text{org}} \) shows a decreasing tendency as the overall distance between CCEs increases. From −6 to 0 hr, while the number of CCEs is still increasing, the newly formed CCEs tend to appear preferentially near existing ones, corresponding to the increasing trend of \( I_{\text{org}} \) (i.e., Phase 1 convective clustering). From around the time of the peak rain rate, as the stratiform clouds begin to develop, two types of CCEs coexist in the domain: (i) relatively aggregated CCEs embedded within stratiform clouds and (ii) relatively isolated CCEs with no clear coherent stratiform structure. For the next 5–10 hr, as the large-scale environment stabilizes, the second type of CCEs dissipate more quickly than the first type. As a result, \( I_{\text{org}} \) increases during the decaying stage from 0 to 10 hr (i.e., Phase 2 convective clustering).

Figures 2e–2h show the corresponding evolution of the simulated reflectivity field in an arbitrarily chosen member of the Control ensemble experiments. Overall, this Control simulation adequately captures the evolution of the number and distribution of CCEs throughout the simulation period, including the transitions from a few localized CCEs (not shown) to scattered convection over a wider region (Figure 2e), to the development of stratiform clouds (Figures 2f and 2g), and to the dissipation of the isolated CCEs (Figure 2h).

Figure 3 compares the time series of rain rate (black), \( N \) (green), and \( I_{\text{org}} \) (red) in the Control ensemble simulations (lower panel; lines represent ensemble mean) with those in observations (upper panel). Consistent with Figure 2, the time evolution of \( I_{\text{org}} \) in the Control shows a steady increase from roughly 10 hr before to 10 hr after the peak rain rate, closely resembling observations. This strongly suggests that the Control
simulations reasonably capture the convective clustering process. The model simulation, however, is not perfect with two main discrepancies. First, the total number of CCEs is lower in Control than in observations, likely due to the horizontal resolution being higher in observations. Second, unlike in observations in which $N$ increases gradually from $-20$ to $0$ hr, $N$ increases rapidly as soon as the model experiences the positive moisture advection forcing (Figure 1d, around $-19$ hr) in the Control simulation. We speculate that this is because all grid points in the domain “feels” the same large-scale forcing that is favorable for convection (Figures 1d and 1e) at the same time. After the rapid increase, $N$ maintains its value around 80 until about $-2$ hr, while the trend of $N$ almost coincides with that of rain rate in observations.

In Figure 3, we marked the time periods that correspond to Phase 1 (light green) and Phase 2 (light red) convective clustering following CKR18 except that $N$ exhibits neither strong negative nor strong positive trend during Phase 1 due to the second discrepancy mentioned above. However, both phases of convective clustering are reasonably well represented in the Control simulations.

### 4.2. Intervention Experiments

In the rest of the section, we examine the intervention experiments to understand the causes of convective clustering shown in Control. $I_{org}$ is used to determine if the targeted field under examination is important for convective clustering: If $I_{org}$ in one of the intervention experiments is significantly lower than $I_{org}$ in Control, it suggests that the mechanism/processes relevant to the targeted fields in the experiment are important for convective clustering.

The time evolutions of $I_{org}$ in each experiment are shown together with those in Control in Figures 4a to 4e. The differences between each experiment and Control are summarized in Figure 4f. A two-tailed Student’s $t$ test is employed to determine the statistical significance of the differences at the 0.01 significant level.

There are two simulations in which $I_{org}$ shows no significant difference from the values of Control: BLWS and RAD. This suggests that the boundary layer wind shear and the inhomogeneity of radiative heating play a minor role in regulating the spatial clustering of convection over this 2-day time scale. Further implications related to these two experiments will be discussed in section 5. On the other hand, $I_{org}$ in BLT, LFTq, and LFTWS are significantly lower than that in Control from $-12$ to $-4$ hr, suggesting that the Phase 1 convective clustering is affected by the processes muted or weakened in those simulations.
Figure 5 shows the radar snapshots from one of the ensemble members of Control (Figure 5a) and of BLT, LFTqv, and LFTWS (Figures 5b and 5c) at $t = -7$ hr, corresponding to the beginning of the Phase 1 period. Figure 5 shows that the distribution of convective clouds is most clearly clustered in Control, where most convective clouds are in the proximity of other convective clouds. In contrast, in the other three experiments, the spatial pattern of convective clouds is shown to be more scattered, consistent with their lower $I_{org}$ (indicated above each panel).

For the Phase 2 period, only LFTqv shows significant difference in $I_{org}$ with a much lower $I_{org}$ than that in Control from 0 to +12 hr. Figure 6 shows the radar snapshots for Control and LFTqv at $t = +5$ hr, corresponding to the end of the Phase 2 period. Convective clouds in Control are embedded in the stratiform area and are mostly localized in a limited region. On the other hand, while the broad stratiform area also appears in LFTqv, the convective clouds are mostly outside the stratiform area, and their distribution is more scattered over the domain when compared to Control.

In the following subsections we will further examine the results of the BLT, LFTqv, and LFTWS experiments. Our analysis will focus on the $-8$ to $-4$ hr and +4 to +8 hr, where $I_{org}$ in these experiments are most significantly different from Control.

### 4.3. BLT

The reduction of $I_{org}$ in BLT (Figure 4a) indicates that the spatial distribution of convection is influenced by the boundary layer temperature perturbations. The reduction is more pronounced in Phase 1 than in Phase 2. Since Phase 1 corresponds to the developing period where convective clouds are continuously forming...
Figure 5. The snapshots of the model reflectivity field and convective clouds from (a) Control simulation, (b) BLT, (c) LFTWS, and (d) LFTqv at $t = -7$ hr, corresponding to the beginning of Phase 1. The gray shading indicates the model reflectivity, and the black shadings indicate the location of the CCEs that are used for calculating $I_{org}$.

Figure 6. Same as Figure 5, but for (a) Control and (b) LFTqv at $t = +5$ hr, corresponding to Phase 2.
(i.e., increasing $N$), this result suggests that the boundary layer temperature perturbation is important for organizing newly triggered convective clouds around the existing ones.

In order to examine the strength of convectively driven cold pools and vertical motion along their edges, we composite temperature and vertical velocity anomalies as a function of the distance to the nearest CCEs ($D_{NC}$). Figure 7a shows maps of $D_{NC}$ from a snapshot in Control at $t = -7$ hr. $D_{NC}$-based composites shown in Figure 7 allow us to examine the difference between convectively clustered and dispersed regions. For example, a region occupied by many CCEs will correspond to a region with small $D_{NC}$, while an area with no CCEs will correspond to a large $D_{NC}$. Figures 8a and 8b show the composited boundary layer temperature perturbations during Phase 1 in both Control and BLT. In Figure 8, the negative boundary layer temperature perturbations near CCEs indicate cold pools around the CCEs. The cold pool strength, measured by the magnitude of the negative temperature anomaly, is gradually decreasing with increasing $D_{NC}$, suggesting that the cold pools are driven by CCEs. The strength of the convectively driven cold pools is stronger in Control than in BLT. The sizes of cold pools, which can be identified as the location of zero temperature perturbation, are also smaller in BLT than in Control.

Figures 8c and 8d show the vertical velocity profiles at the leading edges of the convectively driven cold pools. The positive vertical velocity near the surface in Control suggests that the boundary layer lifting is likely a result of the spreading cold pools. When the lifted air reaches its level of free convection, new convective clouds will be triggered at the leading edge of cold pools (Böing et al., 2012; Khairoutdinov & Randall, 2006; Krueger, 1988). When the boundary layer temperature anomalies are reduced, the cold pool-driven boundary layer lifting becomes much weaker (Figure 8d), further suggesting that spreading cold pools is a major contributor to the boundary layer lifting.

Note that suppressing boundary layer temperature perturbations has a weaker impact on the Phase 2 convective clustering (Figure 4a), suggesting that the cold pool-updraft interaction is less effective of
organizing convection. We suspect that this is probably because in Phase 2, as the stratiform clouds develop, the evaporation of stratiform precipitation generates a broad region of cold pools in the boundary layer. Since stratiform region covers nearly half of the domain (Figure 2g), the domain mean boundary layer temperature is largely cooled as a result. Therefore, the temperature gradient at the leading edge of cold pools becomes weaker (not shown), leading to a weaker lifting force. Thus, the cold pool-updraft interactions become weaker in Phase 2, which explains the smaller difference in $I_{org}$ between Control and BLT in Phase 2.

Lastly, we note that by suppressing the boundary layer temperature perturbations, boundary layer buoyancy is also affected (see supporting information Figure S1). Yang (2018) suggested that the boundary layer buoyancy field is key to convective aggregation in his simulations. While it would take more than 1 day for such boundary layer circulation to fully develop, whether such circulation is important for convective clustering on the time scale of ~1 day is not yet clear and worth further investigation.

### 4.4. LFTWS

It was shown in Figure 4c that $I_{org}$ is significantly lower in LFTWS than in Control, especially in Phase 1. We found that this is because the lower free-tropospheric wind shear has a strong influence on the number of CCEs ($N$) during Phase 1. Figure 9 shows that $N$ in LFTWS is significantly greater than that in Control from $-8$ to $-4$ hr, during which $I_{org}$ in LFTWS is significantly lower than that in Control. The simultaneous increase in $N$ and decrease in $I_{org}$ suggests that the convective clouds can form more easily and are distributed more randomly in LFTWS. This result is consistent with Figure 5, which shows that the convective clouds in LFTWS are more isolated and spatially scattered across the domain than those in Control.

To investigate how the isolated convective clouds can form more easily in LFTWS, we composite the vertical cross section of vertical velocity and horizontal wind of CCEs along the zonal direction, which is the general
Figure 9. The time series of number of CCEs (N) as ensemble mean in Control (black solid line) and in LFTWS (blue solid line). The gray and green shadings indicate the ensemble spread of one standard deviation of the Control and LFTWS, respectively. The green and red shadings indicate the time periods of Phase 1 and Phase 2 in Control, respectively.

Figure 11a shows the profiles of ε in both simulations. For levels between 2 and 4 km, where wind shear is reduced, ε is found to be 20% smaller in LFTWS. The estimated bulk entrainment rate decreases as the difference in εe between updrafts and their environment increases (Figure 11b) and vertical gradient of εe within updrafts decreases (Figure 11c). These results are consistent with the previous studies that developing updrafts entrains a lesser amount of environmental dry air when environmental vertical wind shear is reduced. The above results support the hypothesis that the stronger low free-tropospheric wind shear will enhance the entrainment of environmental dry air into the updrafts. When the vertical wind shear is reduced, the updrafts can develop more easily with a weaker entrainment, resulting in a higher N with a more scattered spatial distribution.

Note that the bulk entrainment rate increases at ~5 km and at ~1.5–2 km. The increase in the bulk entrainment rate at ~5 km is likely because imposing equation (3) in the momentum equations causes a sudden change of horizontal wind near the top of the forced layer, increasing the vertical wind shear, and hence the vertical tilting of cumulus clouds at the upper boundary. At the same time, removing the lower free-tropospheric wind shear would also weaken the boundary layer wind due to the reduced downward momentum flux from above. The weaker boundary layer wind would reduce the surface flux, which would reduce the lower level εe in the updraft pixels (not shown) and lead to a smaller updraft-environment difference of εe (denominator in equation (5)) for levels below 2 km in LFTWS.

4.5. LFTqv

Figure 4b shows that LFTqv is the only experiment whose Iorg is significantly lower than that in Control during both phases of convective clustering. This result suggests that the lower free-tropospheric moisture variation is important for both Phases 1 and 2 convective clustering of CKR18. As discussed in section 3.1, mesoscale moist patches can modulate the spatial distribution of convection, and these moist patches can form through natural fluctuations of moisture (Grabowski & Moncrieff, 2004). To investigate whether the moist patch contributes to the convective clustering, the spatial coherence between convection and moist
patches is examined. Figure 12 shows that CCEs are clustered and located preferentially within the moist patches in Control. When the lower free-tropospheric moisture variation is suppressed in LFTqv, the difference between moist and dry patches is reduced and the convection becomes more scattered. However, while convection is distributed almost randomly when moist and dry patches disappear in Phase 1 (Figure 12b), CCEs in Phase 2 are aggregated even without moist patches (Figure 12d), albeit lower Iorg than in Control. This suggests that processes other than the mesoscale moist and dry patches act to cluster convection in Phase 2.

To examine whether the strength of the mesoscale circulation contributes to convective clustering during Phase 2, we composite the effective stream function ($\psi$) in the height-DNC space. Following Bretherton et al. (2005), $\psi$ is calculated as a horizontal integral of vertical mass flux. It is assumed that the stream function is zero far away from the convection, which will serve as the boundary condition as the integral starts from the right-most bin and integrates leftward on Figure 13. The formulation of $\psi$ can be expressed as follows:

$$\psi(k, i) = \psi(k, i + 1) - \rho(k) \cdot w(k, i + \frac{1}{2})$$

where $k$ indicates the vertical level and $i$ indicates the bin of the distance from the nearest CCE (smaller $i$ corresponds to closer the distance to the nearest CCEs), $\rho$ indicates the density, and $w$ is the bin-averaged vertical velocity.

**Figure 10.** The composite zonal cross-section zonal (arrows) and vertical wind (shadings), and moisture deviations from the domain mean (contours; from 0.02 to 0.05 g/kg) of the CCEs in (a) Control and in (b) LFTWS. The results are composited CCEs with horizontal size smaller than 30 km$^2$ using 5-min output from −8 to −4 hr of five ensemble members.

**Figure 11.** (a) The bulk entrainment rate, (b) the difference in $\theta$ between updraft pixels and environmental air, and (c) $-\partial_\theta \bar{\theta}_e$ profiles for Control and LFTWS. The gray shading indicates the lower free troposphere where the environmental vertical wind shear is reduced in LFTWS.
Figure 12. The snapshots of integrated lower free-tropospheric moisture field from (a) Control and (b) LFTqv at $t = -7$ hr, corresponding to the beginning of Phase 1. The integrated lower free-tropospheric moisture (shading) is calculated by integrating the moisture from 0.8 to 5 km. The black shadings indicate the convective pixels, and the black contours indicate the stratiform regions. (c) and (d) are the same as (a) and (b), but for $t = +5$ hr.

Figure 13. The composite profiles of moisture anomalies (shading), and relative stream function (with 0.005 [m$^2$/s] intervals, all positive values) in (a) Control and in (b) LFTqv. The x axis shows the distance to the nearest CCEs. The corresponding stratiform area fractions in (a) and (b) are shown in (c) and (d). The results are composited using 5-min outputs from +4 to +8 hr using five ensemble members.
The effective stream function in Control during +4 to +8 hr is shown as contours in Figure 13. The contours in the figure are positive with maxima at $z = 5$ km, indicating a clockwise circulation. Figure 13a indicates a mesoscale descending region from $D_{NC} = 20$ to 80 km, where it is largely associated with stratiform area (the corresponding stratiform area fraction is over 40% in both simulations). The dry anomalies in the lower free troposphere in Figure 13a from $D_{NC} = 40$ to 100 km are likely due to unsaturated descent. Note that the spatial relationship is disordered in Figure 13; nonetheless, the overall circulation suggests that the lower-level air ($0$–$3$ km) is transported from the large $D_{NC}$ area to the convective region. Considering that convective heating also contributes significantly to the overall circulation, this suggests that a positive feedback exists between convection, stratiform clouds, and the mesoscale circulation in Control, which becomes much weaker when lower free-tropospheric moisture variation is suppressed.

Figure 13b shows that the relative circulation around convection and stratiform clouds is much weaker in LFTqv, as suggested by the fewer stream function contours. The weakening of the mesoscale circulation in LFTqv is associated with the suppressed mesoscale downdraft in the stratiform region. The mesoscale downdraft is the cold descending air that forms as the stratiform precipitation encounters the relatively dry low-to-middle troposphere air and evaporates. In LFTqv, when the lower free-tropospheric moisture is homogenized, the relatively dry air underneath the stratiform clouds is largely eliminated, hence the weaker evaporative cooling and weaker descending motion. Figure 14 shows the composite diabatic heating profiles in Control and LFTqv during Phase 2. In the low-to-middle free troposphere, both simulations exhibit negative diabatic heating, which is likely due to cooling from evaporation of stratiform precipitation. When the moisture variation is reduced in LFTqv, the dry anomalies underneath the stratiform cloud are reduced, and the contrast in diabatic heating between convective and stratiform regions is weakened. This result explains the weakening of the mesoscale downdraft in LFTqv (Figure 14b). As the strength of the mesoscale circulation decreases in LFTqv, convection becomes more scattered (Figures 4b and 12d).

In short, our results suggest that lower free-tropospheric moisture variation affects convective clustering at least in two ways: (i) In Phase 1, convection forms preferentially within mesoscale moist patches; and (ii) in Phase 2, dry anomalies underneath stratiform clouds strengthen the stratiform-driven mesoscale downdrafts that help localize convection in the convective region.

**5. Discussion**

a. Noncritical mechanisms for convective clustering

We found that the boundary layer wind shear is not a critical mechanism for convective clustering during the 2-day rain event in our study. While the interaction between cold pools and the background vertical wind shear in the boundary layer may be important for maintaining convection in a localized region over midlatitude continental regions (Keenan & Carbone, 1992; Robe & Emanuel, 2001; Bryan et al., 2006), the extent to which it can be applied to tropical oceanic convection has been questioned in recent studies (Anber et al., 2014, 2016; Grant et al., 2018; Lane & Moncrieff, 2015; Moncrieff & Lane, 2015). Our result also suggests that the RKW theory is not responsible for the convective clustering in this observed case. A possible explanation...
is that the buoyancy and depth of the cold pools is weaker and shallower over the ocean compared to over land; therefore, the balance between cold pools and the background wind shear plays a minor role in maintaining the aggregated convection.

While radiative-driven circulations are suggested to be important for triggering and maintaining convective self-aggregation in idealized studies (Bretherton et al., 2005; Muller & Bony, 2015; Wing & Emanuel, 2014), our results show no role of this radiative impact on convective clustering in an observed 2-day rain event simulation. This difference can be explained by the two key differences between our simulation and the idealized RCE simulations: (i) Our simulation is forced by large-scale forcing, which means that the simulation is not in RCE and may therefore have different clustering mechanisms; and (ii) the convective clustering process in our simulation has a shorter time scale of ~1 day, which may not be long enough for the radiative-forcing-driven circulation to develop.

b. Key mechanisms for convective clustering

Our results show that the interaction of boundary layer cold pools with convective updrafts is one of the essential processes for convective clustering during Phase 1. The spreading boundary layer cold pools are found to be responsible for triggering new updrafts near the existing convective clouds by providing additional lifting at the leading edge. This result agrees with previous studies that highlight the lifting mechanism by cold pools (Böing et al., 2012; Khairoutdinov & Randall, 2006). However, it is found in our study that the effect of this mechanism on convective clustering is more pronounced in Phase 1 than in Phase 2. This result might be due to the development of stratiform cloud in Phase 2. As the broad stratiform clouds form, the boundary layer temperature is cooled over a large portion of the domain as a result of the evaporation of stratiform precipitation and the mesoscale downdraft, making the temperature gradient at the leading edge of cold pools weaker. This weakens the lifting force of the spreading cold pools and diminishes the effect of cold pools on convective clustering in Phase 2.

Our results also show that the lower free-tropospheric wind shear is a critical factor for convective clustering, and the effect is more pronounced during Phase 1 than Phase 2. We find that, with a stronger lower free-tropospheric wind shear, the entrainment of environmental air into updrafts increases, which is consistent with previous studies (Malkus, 1952; Scorer & Ludlam, 1953). This increase in the entrainment constrains the convection to form preferentially within relatively moist regions. In Phase 1, the environment is unstable and is favorable to convection. The environmental wind shear at this time acts as an effective filter, preventing convection from being triggered in dry regions. When entering Phase 2, this effect of environmental wind shear on convective development becomes weaker likely because the environment becomes more stable and the development of convection often requires stronger forcing, such as boundary layer convergence due to the mesoscale circulation.

The lower free-tropospheric moisture variation is found to be a key ingredient for convective clustering during both Phases 1 and 2 but may be through different mechanisms. Convection tends to form preferentially in the region with higher lower free-tropospheric moisture, which localizes the convection within the mesoscale moist patches in Phase 1. Therefore, the natural fluctuation of the lower free-tropospheric moisture may help to constrain the convection to a certain region of the domain, leading to convective clustering. Colin et al. (2019) drew a similar conclusion from a set of CRM experiments. In their study, they examined the response of a wind shear-driven mesoscale organized convective system to a sudden moisture homogenization. They found that the mesoscale moisture fluctuations provide the dominant memory to the organized convective system, emphasizing the role of moisture variation.

In Phase 2, the lower free-tropospheric moisture variation controls the strength of the mesoscale circulation associated with the development of stratiform clouds. As the convective systems mature and stratiform clouds develop, diabatic heating in the convective region and the evaporative cooling in the stratiform region develop a mesoscale circulation that enhance the boundary layer convergence in the convective region and help localize convection. This positive feedback between convective and stratiform clouds and the mesoscale circulation weakens significantly when lower free-tropospheric moisture variation is suppressed, due to weaker evaporative cooling and a weaker mesoscale downdraft in the stratiform region (Virman et al., 2018), resulting in a more scattered distribution of convection. It is important to note that the mesoscale downdraft is not unique to squall-line-type convective systems but also observed in other types of MCSs (e.g., Kingsmill & Houze, 1999; Smull & Houze, 1987; Zipser & Gautier, 1978). In an observational study
during AMIE/DYNAMO, Barnes and Houze (2014) documented that the leading-convective-line/trailing-stratiform structure was present in most December rain events in AMIE/DYNAMO, but it was absent in most October and November events. Despite the differences in the morphology of convective systems, it was documented that the presence of descending midlevel inflow layer was a common feature in these rain events.

6. Summary and Conclusions

The current study investigates the mechanisms of the observed convective clustering in a 2-day rain event during the AMIE/DYNAMO field campaign. By inspecting the trends in the number of convective clouds (N) and the degree of convective clustering (Iorg), CKR18 classified the convective clustering processes into an intensifying stage (Phase 1; increasing N and Iorg) and a decaying stage (Phase 2; decreasing N and increasing Iorg). The goal of this study is to understand the mechanisms through which convection becomes clustered during these two different phases.

The mechanisms of convective clustering are investigated by analyzing a series of ensemble simulations performed using the WRF model. In our control ensemble simulations, the observed convective clustering is realistically captured in terms of the trends of N and Iorg. A set of intervention experiments is designed to test the possible mechanisms of convective clustering. All intervention simulations and the targeted mechanisms are briefly summarized in Table 1.

The experiment results are examined to infer the cause of convective clustering in Control. The results from BLWS and RAD experiments show that the convection is continuously clustering over time without the presence of boundary layer vertical wind shear and horizontal inhomogeneity of radiative heating, suggesting that mechanisms related to these two fields play a minor role in the observed convective clustering during the 2-day rain events. On the other hand, the experiment results from BLT and LFTqv show that the cold pool-updraft interactions, moisture-convection feedbacks, and mesoscale circulations are all important factors for convective clustering. While these results may not be surprising because they agree with the findings from midlatitude squall-line systems, it is still important to confirm that these mechanisms are applicable to the clustering of tropical oceanic convection, even for non-squall-line convective systems.

An important contribution of this study is that we show that the background lower free-tropospheric wind shear can control the degree to which convective clouds aggregate, especially in the developing phase. As the convective clouds preferentially form in the moist patch, the lower free-tropospheric wind shear acts as another driving force to further prevent convection from triggering in the drier region, enhancing convective clustering. In the later stage (Phase 2), as the convective system matures, the importance of cold pools and vertical wind shear weaken, as convection is able to cluster without the boundary layer temperature perturbation and lower free-tropospheric wind shear. Instead, the mesoscale circulation associated with the developing stratiform clouds is a key ingredient for convective clustering in this later stage.

Our results suggest that the stratiform clouds can affect convective clustering at least through the following three effects: (i) The boundary layer is cooled over a wide region, which weakens the cold pool-updraft interaction; (ii) the boundary layer cooling over a wide region suggests that the convection can only develop in a smaller domain that is not covered by stratiform clouds; and (iii) the mesoscale downdraft associated with stratiform clouds helps form a mesoscale circulation, which localizes the convection. Among the three effects, (i) hinders the convective clustering, while (ii) and (iii) aid convective clustering. Our results suggest that, in Phase 2 of the 2-day rain event, the combined effect of (ii) and (iii) outweighs the effect of (i). However, the relative importance of these effects is worthy of a further investigation.

Our study provides guidance for the parameterization of convective organization for tropical oceanic convection. Some recent convection schemes have attempted to incorporate the representation of convective clustering. However, most of them focus on the cold pool-updraft interaction processes (see table 1 in CKR18 and the references therein). While this study confirms that the cold pool-updraft interaction is indeed an important contributor for convective clustering during Phase 1, it also suggests that additional mechanisms are critical for convective clustering in tropical oceanic convective systems on the time scale of 1 day. In particular, the lower free-tropospheric vertical wind shear, subgrid-scale moisture variations in
the lower free troposphere, stratiform clouds, and the associated mesoscale circulation should also be considered in the future development of parameterization of convective organization.

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