Probing the Response of Tropical Deep Convection to Aerosol Perturbations Using Idealized Cloud-Resolving Simulations with Parameterized Large-Scale Dynamics

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ABSTRACT

A framework is introduced to investigate the indirect effect of aerosol loading on tropical deep convection using three-dimensional limited-domain idealized cloud-system-resolving model simulations coupled with large-scale dynamics over fixed sea surface temperature. The large-scale circulation is parameterized using the spectral weak temperature gradient (WTG) approximation that utilizes the dominant balance between adiabatic cooling and diabatic heating in the tropics. The aerosol loading effect is examined by varying the number of cloud condensation nuclei (CCN) available to form cloud droplets in the two-moment bulk microphysics scheme over a wide range of environments from 30 to 5000 cm⁻³. The radiative heating is held at a constant prescribed rate in order to isolate the microphysical effects. Analyses are performed over the period after equilibrium is achieved between convection and the large-scale environment. Mean precipitation is found to decrease modestly and monotonically when the aerosol number concentration increases as convection gets weaker, despite the increase in cloud liquid water in the warm-rain region and ice crystals aloft. This reduction is traced down to the reduction in surface enthalpy fluxes as an energy source to the atmospheric column induced by the coupling of the large-scale motion, though the gross moist stability remains constant. Increasing CCN concentration leads to 1) a cooler free troposphere because of a reduction in the diabatic heating and 2) a warmer boundary layer because of suppressed evaporative cooling. This dipole temperature structure is associated with anomalously descending large-scale vertical motion above the boundary layer and ascending motion at lower levels. Sensitivity tests suggest that changes in convection and mean precipitation are unlikely to be caused by the impact of aerosols on cloud droplets and microphysical properties but rather by accounting for the feedback from convective adjustment with the large-scale dynamics. Furthermore, a simple scaling argument is derived based on the vertically integrated moist static energy budget, which enables estimation of changes in precipitation given known changes in surface enthalpy fluxes and the constant gross moist stability. The impact on cloud hydrometeors and microphysical properties is also examined, and it is consistent with the macrophysical picture.

1. Introduction

Atmospheric aerosol particles may act as cloud condensation nuclei (CCN) over which cloud droplets form,

and therefore have important microphysical effects. Increasing concentration of aerosols changes the structure and composition of clouds, and modifies convection (e.g., Tao et al. 2012; Fan et al. 2016). Such cloud adjustments due to aerosols have a large impact on surface precipitation and its forming processes (e.g., IPCC 2014).

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These complex interactions take place over multiscale flows, which involve microphysical, thermodynamical, and dynamical feedbacks over relatively longer time scale than a single case event with short cloud–aerosol response time scale.

The simplest approach to investigate these feedbacks is one that grants a balance between radiative cooling and convective heating rate [radiative–convective equilibrium (RCE); e.g., Emanuel 2007]. This choice implies no domain-mean vertical motion W . For example, Grabowski (2006) and Grabowski and Morrison (2011) used the RCE framework over fixed sea surface temperature (SST) to mimic Earth's mean conditions and found that surface precipitation remains almost unchanged in response to CCN perturbations because of the fact that precipitation is constrained almost entirely by radiative cooling of the atmosphere. Van den Heever et al. (2011) and Storer and van den Heever (2013) also used the RCE approach but with a large domain to allow spontaneous circulations to develop as a result of organized convection. They found that increasing aerosols concentration tends to enhanced convective mass flux and associated precipitation in the moist regions, although the effect on domain-mean precipitation remains small.

Another approach to examine the aerosol–convection effects over long time scale is one in which the large-scale forcings (e.g., W) are prescribed from observational datasets. For instance, Seifert et al. (2012) conducted simulations of a 3-yr climatology of aerosol effects on summertime precipitation over Germany using a high-resolution nonhydrostatic model and a two-moment microphysics scheme. They showed that high CCN concentration decreases warm-rain formation (autoconversion) because of the smaller cloud droplets, which is accompanied by an increase in liquid water content and convective invigoration due to additional latent heat release (Seifert and Beheng 2006; Fan et al. 2013, 2018). The effects of CCN on aerial-averaged accumulated surface precipitation are small, confirming the role of clouds acting as a buffered system (e.g., Grabowski et al. 1998; Stevens and Feingold 2009).

Invigoration of convective clouds was also found by Fan et al. (2013), who performed a month-long cloud-resolving model simulations with spectral bin microphysics of tropical summer clouds over the tropics and extratropics forced by observations from western tropical Pacific and southern Great Plains datasets, respectively. Early-stage cloud invigoration was found to be induced by additional latent heat release from cloud liquid water content and water freezing at higher altitudes. However, at mature cloud stages, invigoration was found to be dominated by the aerosol-microphysical effects in the polluted environment leading to smaller and longer-lasting ice

particles in the anvil clouds. Rain rate was found to be slightly higher in polluted environments.

A major drawback in the above studies is that surface precipitation is strongly constrained by the observationally derived large-scale forcing, as noted by Morrison and Grabowski (2011). They found that increasing aerosols has a small impact on precipitation and leads to a weaker convection because of an additional feedback of convection–large-scale circulation adjustment.

In reality, in the tropical atmosphere where horizontal gradients of temperature and density are weak, small changes in convective heating induce small (but significant) deviations in atmospheric temperatures. The resulting large-scale vertical motion W then acts in a way to restore tropospheric temperature and maintain its horizontal homogeneity over large variations in surface forcings (e.g., Bretherton and Smolarkiewicz 1989). This mechanism of a two-way interaction between convection and large-scale motion has long been implemented in single-column models (e.g., Sobel and Bretherton 2000; Ramsay and Sobel 2011) as well as cloud-resolving models (e.g., Raymond and Zeng 2005; Wang and Sobel 2011; Romps 2012; Daleu et al. 2012; Anber et al. 2014, 2015a,b). Varying SST while strongly relaxing free-tropospheric temperature toward a fixed profile can be thought of as varying the relative SST, or the difference between local and tropical mean conditions.

In our study here, we use this approach of higher relative SST to force deep convection in a 3D cloud-system-resolving model (CRM) that uses weak temperature gradient (WTG) approximation to diagnose W . We then vary the aerosol number concentration over a wide range of environments from pristine to very polluted. Introducing this additional feedback between convective clouds and the large-scale dynamics at quasi equilibrium warrants a more realistic assessment of the effect of aerosols perturbations on precipitation and convection as opposed to bubble experiments.

Furthermore, parameterizing W in the domain would mimic a scenario of a small tropical region experiencing deep convection. Under this scenario, and unlike the conventional method in which W is prescribed, both convection and the large-scale dynamics interact with each other. This allows a superior evaluation of changes in the domain-mean precipitation that is no longer tied to radiative heating alone as in RCE, which would represent the mean state of the entire tropics, neglecting the energy transport by the large-scale circulation.

This paper is organized as follows. In section 2 we describe the parameterization of the large-scale circulation in the limited domain CRMs, the model setup, and the experiment design. In section 3, we show how mean precipitation responds to varying aerosol number

concentration and present a simple scaling argument based on conserved variables of moist and dry static energy to explain this response. We present bulk properties of cloud hydrometeors and number concentration profiles in [section 4](#). In [section 5](#), we perform a number of sensitivity tests, and summarize in [section 6](#).

2. Model description and experimental design

a. Numerical model

We use the Advanced Research Weather Research and Forecasting (WRF) Model, version 3.5 ([Skamarock et al. 2008](#)). We adopt the following physics schemes:

- Boundary layer turbulence and vertical subgrid eddies are parameterized using the Yonsei University (YSU) scheme ([Hong and Pan 1996](#); [Noh et al. 2003](#); [Hong et al. 2006](#)).
- Horizontal subgrid eddies are treated using a Smagorinsky first-order closure ([Xue et al. 2000](#); [Skamarock et al. 2008](#)).
- Surface moisture and heat fluxes are parameterized following Monin–Obukhov similarity theory ([Monin and Obukhov 1954](#)).
- The horizontal and vertical advection schemes are fifth and third order, respectively ([Skamarock et al. 2008](#)).
- Moisture and other scalars are advected using a positive-definite scheme ([Skamarock et al. 2008](#)).
- We use the implicit scheme ([Klemp et al. 2008](#)) to damp unphysical reflection of vertically propagating gravity waves in the top 5 km of the numerical grid.
- A simplified radiative heating is applied in place of interactive radiation at a constant and uniform rate of 1.5 K day^{-1} in the troposphere (defined below), which is close to the observed climatology, so that cloud–radiative feedback is not included.
- The Morrison scheme ([Morrison et al. 2005, 2009](#)) is used for cloud microphysics, which is a double-moment bulk scheme that predicts mixing ratios and number concentrations of five hydrometeor species: cloud droplets, ice, rain, snow, and graupel.

All the numerical experiments were performed with horizontal grid spacing of 2 km and constant Coriolis parameter of $f = 0$. We use 50 vertical levels with 10 levels in the lowest 1 km and a grid spacing gradually increasing to 1.5 km near the model top at ~ 22 km. The computational domain has 96×96 horizontal grid points and doubly periodic lateral boundary conditions.

b. Parameterized large-scale dynamics

We use the spectral WTG method introduced by [Herman and Raymond \(2014\)](#) to parameterize large-scale

dynamics based on modification by [Wang et al. \(2016\)](#). In the tropical WTG environment, temperature anomalies induced by convection are quickly removed by means of gravity wave adjustment resulting in a balance between the diabatic heating and adiabatic cooling in the thermodynamic equation. Since different heating profiles result in gravity waves of different phase speeds, we project heating anomalies onto a set of orthogonal spectral modes based on the vertical structure equation of large-scale vertical velocity W . The value of W is derived from the RCE sounding, as described in [Wang et al. \(2016\)](#). Specifically, we expand W as

$$W = \sum_{n=1}^N \tau_1^{-1} \left(\frac{c_n}{c_1} \right) j_n W_n \quad (1)$$

where τ_1 and c_1 are the time scale and phase speed of the gravest mode, respectively; c_n is the phase speed of the gravity wave mode n with vertical structure as W_n , and j_n is the projection of potential temperature anomaly onto gravity wave mode n :

$$j_n = \left\langle \frac{\overline{\theta}_v - \theta_v^B}{\partial_z \overline{\theta}_v}, W_n \right\rangle \left| \langle W_n, W_n \rangle \right|^{-1/2}, \quad (2)$$

where $\overline{\theta}_v(z, t)$ is the domain-averaged virtual potential temperature resulting from the model simulated convective heating, and $\theta_v^B(z)$ is the domain-averaged virtual potential temperature from the RCE run ([Fig. 1](#)). The inner product between two quantities as in [Eq. \(2\)](#) is defined as

$$\langle W_m, W_n \rangle = \int_0^H W_m W_n N^2 dz = \langle W_n, W_n \rangle \delta_{mn}, \quad (3)$$

where $N^2 = (g/\theta_v)(\partial\theta_v/\partial z)$ is the Brunt–Väissälä frequency, θ_v is the environment potential temperature, g is the gravitational acceleration, and δ_{mn} is the Kronecker delta. Here, we set $\tau_1 = 2$ h, which is interpreted as the time scale for the first gravity wave mode to propagate out of the domain. For our domain length of 192 km, this gives gravest mode phase speed of $c_1 \approx 25 \text{ m s}^{-1}$.

Parameterized large-scale vertical motion W obtained in [Eq. \(1\)](#) above introduces additional sources and sinks to thermodynamic and dynamic variables. Its effect on dynamical variables (e.g., momentum) is very small by scaling arguments and is thus omitted ([Herman and Raymond 2014](#)). In this work, we consider the effect of W on two variables: temperature and water vapor. Adiabatic heating/cooling and vertical transport of moisture may be written as an additional term associated with W in the potential temperature and moisture equations, respectively, as follows:

FIG. 1. Vertical profiles of (a) potential temperature and (b) water vapor mixing ratio for the RCE run at equilibrium, which are used to initialize WTG experiments.

$$\left(\frac{\partial \theta}{\partial t}\right)_{\text{LS}} = \dots - W_{\text{WTG}} \frac{\partial \bar{\theta}}{\partial z}, \quad (4)$$

$$\left(\frac{\partial q}{\partial t}\right)_{\text{LS}} = \dots - W_{\text{WTG}} \frac{\partial \bar{q}}{\partial z}, \quad (5)$$

where θ and q are potential temperature and specific humidity of water vapor, respectively, and the subscript “LS” denotes large scale. Note that, unlike the conventional WTG (e.g., Daleu et al. 2017) this method does not require a special treatment in the well-mixed boundary layer (e.g., Sobel and Bretherton 2000), as vertical gradients vanish, because the shallow modes with small phase speed are weakly relaxed toward the target profile.

c. Experimental design

Observations indicate that the CCN concentration vary from about 10 cm^{-3} in pristine conditions to 10^3 cm^{-3} in highly polluted environments over the tropical oceans (Fridlind et al. 2012). We use a prescribed number concentration of CCN in the Morrison microphysics scheme as the only control parameter in our study. Concentration of activated cloud droplets $N_{\text{ccn}} (\text{cm}^{-3})$ is computed as a function of the supersaturation $S = (q_v/q_s - 1) \times 100\%$, assuming a specified number of CCN, through a power-law relationship: $N_{\text{ccn}} = C_N S^k$, where C_N is the prescribed concentration of CCN, which is activated at a critical supersaturation value of $S = 1\%$, and k is an empirical constant taken to be 0.4, a typical value for maritime conditions (Pruppacher and Klett 1997; Rasmussen et al. 2002;

Khairoutdinov and Yang 2013). Prognostic aerosols are not considered in the present study, and changes in the aerosol size and chemical composition due to cloud–aerosol interactions are neglected. We only consider a simplified approach to these processes by changing the constant parameter C_N as a proxy for changes in aerosol burden, which remains fixed during the simulations.

All experiments performed here have a prescribed noninteractive radiative heating profile to disconnect the direct from indirect aerosols effects. Radiation is a simple Newtonian relaxation scheme as in Pauluis and Garner (2006), Wang and Sobel (2011), and Anber et al. (2014):

$$Q_R = \begin{cases} -1.5 \text{ K day}^{-1}, & T > 207.5 \text{ K} \\ \frac{200 \text{ K} - T}{5 \text{ days}}, & \text{otherwise} \end{cases}. \quad (6)$$

The total column integrated radiative heating is 122 W m^{-2} .

We first perform an RCE experiment at SST of 28°C , with CCN concentration of 200 cm^{-3} until equilibrium is reached after about 60 days. Results from this experiment are averaged over the last 20 days after equilibrium (60–80 days) to obtain statistically steady vertical profiles of potential temperature and water vapor (Figs. 1a and 1b, respectively), representing tropical mean conditions. These RCE profiles are then used to initial conditions as well as target profiles representing the tropical mean conditions in the WTG experiments. The simulated

FIG. 2. Vertical profiles of domain-mean (a) large-scale vertical velocity, (b) potential temperature, and (c) moist static energy for the control WTG experiment with SST = 30°C and $C_N = 200 \text{ cm}^{-3}$.

potential temperature profiles are relaxed toward the target potential temperature profile using Eq. (2).

WTG experiments are performed at SST = 30°C (2 K warmer than the RCE) in order to generate a state of ascending motion mimicking a deep moist convective tropical region (e.g., Ramsay and Sobel 2011; Wang and Sobel 2011, 2012). In the WTG runs, we set C_N to the following values: 30, 50, 100, 200, 500, 1000, 2000, and 5000 cm^{-3} to represent a wide range of environments as we noted above. All WTG experiments are run for 60 days and the averages are made over the last 20 days.

WTG experiment with $C_N = 200 \text{ cm}^{-3}$ is our designated control run against which we compare the effect of varying aerosol burden. Profiles of the mean large-scale vertical velocity W_{WTG} , potential temperature (or roughly the dry static energy), and the moist static energy (MSE) (sum of thermal, latent, and geopotential energy, which is approximately a conserved quantity) of the control run are shown in Fig. 2. Vertical motion (Fig. 2a) is top-heavy, exceeding 15 cm s^{-1} in the upper troposphere at about 350 hPa, reflecting the large energy departure from the RCE state induced by 2-K surface warming advected upward to stabilize the atmosphere and retain the moist adiabat (Figs. 2b and 2c, respectively).

3. Results

a. Large-scale circulation and thermodynamic structure

We first investigate how the large-scale vertical motion and thermodynamic variables respond to

varying CCN concentration. Figure 3a shows the domain-averaged large-scale vertical velocity W_{WTG} (as deviation from the control case at $C_N = 200 \text{ cm}^{-3}$), which exhibits a dipole structure: decreasing from ascending to descending above 800 hPa as CCN concentration increases, indicating more export of MSE (Fig. 3c). Note that in the boundary layer (BL), temperature relaxation is not applicable because most of the transport is done by turbulence and not waves and the temperature adjusts directly to the underlying surface. In the free troposphere, variations in W_{WTG} are coupled to variations in the simulated diabatic heating induced only by latent heat release.

In the troposphere, increasing aerosol number concentration is accompanied by less diabatic heating (Fig. 3e), anomalously descending large-scale motion, anomalously cooler temperatures (Fig. 3b), and less water vapor (Fig. 3d). Below 850 hPa, the small increase in atmospheric temperature is associated with a drier and warmer BL and reduced evaporative cooling resulting from evaporation of falling precipitation as in Fig. 3f. This, in turn, decreases the air–sea temperature disequilibrium and reduces surface enthalpy fluxes as discussed below (see Fig. 4b).

b. Precipitation and energy budget analysis

Time- and domain-mean precipitation decreases monotonically as a function of the added CCN concentration as shown in Fig. 4a. To understand the controls on precipitation in more details we analyze the dry and moist static energy budget.

FIG. 3. Domain-mean vertical profiles of (a) large-scale vertical velocity, (b) temperature, (c) moist static energy, (d) water vapor mixing ratio, (e) diabatic heating, and (f) evaporation from precipitation as a departure (simulation minus control) from the control case profiles ($C_n = 200 \text{ cm}^{-3}$). Results are shown for different aerosol number concentrations.

In quasi-equilibrium state the vertically integrated dry and moist static energy can be written, respectively, as

$$\overline{W} \frac{\partial \bar{s}}{\partial z} = H + P + Q_R, \quad (7)$$

$$\overline{W} \frac{\partial \bar{h}}{\partial z} = E + H + Q_R, \quad (8)$$

where $\overline{W}$ is the large-scale vertical velocity, $s = c_p T + gz$ is the dry static energy, and $h = c_p T + L_v q + gz$ is the moist static energy. Here, c_p , L_v , and g are the specific heat, latent heat of vaporization, and gravitational acceleration, respectively. The T , q , and z are temperature, water vapor mixing ratio, and height, respectively; H , E , and P are sensible heat flux, latent heat flux, and precipitation at the surface, respectively; Q_R is the column-integrated radiative heating.

By combining Eqs. (7) and (8) above, we get a single diagnostic equation of precipitation:

$$P = \frac{1}{M} (H + E + Q_R) - Q_R - H, \quad (9)$$

where M is the normalized gross moist stability, a dimensionless number estimated as

$$M = \frac{\overline{W} \frac{\partial \bar{h}}{\partial z}}{\overline{W} \frac{\partial \bar{s}}{\partial z}}. \quad (10)$$

The gross moist stability is defined as the rate at which the circulation exports moist static energy from a column for a given rate of dry static energy export (or moisture import to the column) (e.g., Neelin and Held 1987; Sobel 2007; Raymond et al. 2007, 2009; Anber et al. 2016).

It is worth mentioning that in Eq. (9), precipitation is not an explicit function of microphysical processes, but rather a function of large-scale forcings: column energy sources (surface enthalpy fluxes, and radiative heating) and the export of column energy by large-scale vertical motion M . Equation (9) also suggests that when M is constant, changes in precipitation result directly from changes in the energy sources of surface enthalpy fluxes

FIG. 4. Mean values of (a) surface precipitation, (b) surface enthalpy fluxes (sum of sensible and latent heat fluxes), (c) sensible heat flux, and (d) latent heat flux for different aerosol number concentration. The values of surface precipitation are those from the model output (blue) and those calculated from the budget equation [Eq. (9), red].

and/or radiative fluxes in the atmospheric column and that the large-scale vertical motion maintains the same vertical shape (or same convective regime). When M varies, however, changes in precipitation can be attributed partly and indirectly to changes in the shape of the large-scale vertical motion, which acts to either add or remove energy from the column (Anber et al. 2015a,b, 2016). Generally, there is no accepted theory that satisfactorily predicts the gross moist stability since it is a function of both state and flow variables.

Fortunately, Eq. (10) above yields a nearly constant M for all the simulations varying CCN concentrations, and is equal to ~ 0.42 . This means that increasing CCN concentration does not change the vertical structure of the large-scale vertical motion and that the level of the top-heaviness of W remains the same as in Fig. 2a. In other words, the normalized large-scale vertical velocities $[W/\max(W)]$ for all the CCN concentration cases coincide vertically (not shown), exporting the same amount of moist static energy.

Taking the partial derivatives of Eq. (8) with respect to logarithmic scale of aerosol number concentration gives

$$\frac{\partial P}{\partial \log_{10}(\text{CCN})} = \frac{1}{M} \frac{\partial(E+H)}{\partial \log_{10}(\text{CCN})} - \frac{\partial H}{\partial \log_{10}(\text{CCN})}, \quad (11)$$

where changes in radiative heating with CCN concentration are zero, since it is held fixed in these simulations [see Eq. (6)].

Facilitated by a constant M , Eq. (11) has a predictive power and suggests that changes in precipitation with CCN concentrations result from changes in surface enthalpy fluxes, which also exhibit a monotonic decrease at same rate of precipitation, as shown in Fig. 4. That is, given M and the magnitude of surface fluxes, changes in precipitation can be estimated directly as a function of aerosol number concentration (see Fig. 4a).

4. Cloud hydrometeors statistics

Figure 5 shows the vertical profiles of mixing ratio of nonprecipitable water (cloud liquid water and ice in Figs. 5a and 5b, respectively), and precipitable water (snow, rain, and graupel in Figs. 5c–e, respectively), as departure from the control run profiles at $C_N = 200 \text{ cm}^{-3}$. Increasing CCN concentration increases the cloud liquid water mixing ratio in the free-tropospheric warm-rain region above 800 hPa, and also increases cloud ice in the upper levels (Fig. 5b). This result is primarily due to the suppression of rain and snow production (Figs. 5d and 5e) resulting from reduced collision efficiencies of smaller droplets (e.g., Tao et al. 2012). Graupel, however, exhibits

FIG. 5. Domain-mean vertical profiles of mixing ratio of (a) liquid water, (b) ice, (c) snow, (d) rain, and (e) graupel as a departure from the control case profiles ($C_n = 200 \text{ cm}^{-3}$). Results are shown for different aerosol number concentrations.

an opposite trend (Fig. 5e). Below 800 hPa, reduction in liquid water mixing ratio might be due to evaporation of smaller droplets caused by entrainment drying (e.g., Xue and Feingold 2006).

Figure 6 shows the corresponding vertical profiles of the number concentration of liquid water, ice, snow, rain, and graupel (Figs. 6a–e, respectively). As with the mixing ratios, increasing CCN concentration monotonically increases the number of liquid water particle number concentration while decreasing their sizes, particularly at about 900 and 600 hPa (Fig. 6a), and increases the ice particle number concentration in the upper troposphere (Fig. 6b). Rain and snow number concentrations (Figs. 6d and 6c, respectively) also decrease, while graupel shows opposite sign (Fig. 6e).

Overall the microphysical picture is consistent with the macrophysical behavior of decreasing surface precipitation. However, as we will discuss below, despite enhanced cloud water mixing ratio in the troposphere, weaker convection and precipitation in response to increasing CCN concentration result mainly because of weaker forcing from surface fluxes by coupling convection

with the large-scale dynamics and not by microphysical considerations such as changes in particle droplet size and concentration.

5. Feedback sensitivity tests

In this section, we focus on the importance of the feedbacks from surface enthalpy fluxes and large-scale vertical motion, as well as the effect of vertical wind shear.

a. Surface fluxes feedback

We performed a set of simulations with three CCN concentrations (30, 200, and 5000 cm^{-3}) but with prescribed (fixed) surface fluxes from the control case at $\text{CCN} = 200 \text{ cm}^{-3}$ so that feedback from surface fluxes is entirely eliminated. Surface precipitation now remains almost constant as aerosol concentration increases above the control value, with a reduction below that (in which case M slightly increases) as shown in Fig. 7a, despite having similar microphysics hydrometeor profiles of mixing ratio and number concentration as in the case of interactive surface fluxes (i.e., similar to Fig. 5).

FIG. 6. As in Fig. 5, but for number concentration of (a) droplet liquid water ($\times 10^2$), (b) ice ($\times 10^1$), (c) snow ($\times 10^{-3}$), (d) rain ($\times 10^{-3}$), and (e) graupel ($\times 10^{-4}$).

This demonstrates the role of surface fluxes feedback in evaluating changes in precipitation, and that microphysical processes alone are insufficient to explain precipitation efficiency in this regard. Our results here agree with those discussed by Morrison and Grabowski (2013) and question the notion of convective invigoration proposed by several studies (e.g., Rosenfeld et al. 2008; Fan et al. 2013, 2018). One possible explanation is that those studies use prescribed large-scale forcings derived from observations and do not account for adjustment of convection with the environment. This adjustment allows the strength of convective dynamics to be determined by cloud buoyancy anomalies and their balance with the environmental cooling/large-scale vertical motion (Robe and Emanuel 1996), illustrating that the effect of droplet size and concentration changes due to aerosols perturbations are insignificant (e.g., Grabowski and Morrison 2016; Grabowski and Jarecka 2015). This is further illustrated below when we fix the large-scale vertical motion.

b. Interactive W_{WTG} feedback

It follows straightforward that in RCE precipitation remains unchanged under aerosols perturbations

because of constrained surface evaporation by radiative heating that is held fixed (Fig. 7b) (e.g., Morrison and Grabowski 2011).

While an RCE experiment does not have a large-scale vertical motion by construction, parameterizing W with WTG approximation in a way that both local and tropical mean SSTs are taken to be the same (i.e., 28°C) can be thought of as an RCE that is susceptible of developing circulation if strong perturbations are introduced. Convective regimes are then determined by the vertical structure of the developed W (bottom-heavy or top-heavy, for example) or the gross moist stability (e.g., Raymond and Sessions 2007; Wang and Sobel 2012; Anber et al. 2015a), and the resulting precipitation is said to be dynamically driven. In this case, perturbations in aerosol number concentration are unlikely to affect the stability of tropical atmosphere and convection as shown in Fig. 7c, which explains why M remains constant (see section 3b).

When the large-scale feedback is disabled, as we do by fixing W_{WTG} from the control run in Fig. 2a, surface precipitation almost remains unchanged, as shown in Fig. 7d, because of the strong control of the large-scale vertical motion (e.g., Morrison and Grabowski 2011;

FIG. 7. Sensitivity of precipitation response as a function of CCN concentration in experiments with (a) fixed surface fluxes, (b) RCE, (c) parameterized W_{WTG} at SST = 28°C, and (d) fixed large-scale W at SST = 30°C.

Lee et al. 2008). These results further demonstrate the insignificance of aerosols-microphysical interactions alone on changing convection and precipitation.

c. Vertical wind shear

Finally, we performed simulations with imposed deep and strong vertical wind shear in which mean winds are strongly relaxed to a constant wind profile extending from the surface at speed of 5 m s^{-1} to the depth of the troposphere at 35 m s^{-1} . The purpose of the shear is to force convective organization into squall-line-like structure (Anber et al. 2014 and their Fig. 2). Mean precipitation in this case exhibits similar behavior to the simulations without shear (not shown), which is diagnosed by Eq. (9) (e.g., Anber et al. 2014).

6. Summary and conclusions

We have investigated the role aerosol effects play in modifying convection and rainfall rate using a series of

idealized limited domain cloud-system-resolving model simulations with parameterized large-scale circulation, representing a tropical region over convective SST above the tropical mean. The model operates on a doubly periodic domain and the large-scale dynamics is parameterized using the spectral WTG approximation method in which the adiabatic cooling has a spectral decomposition in the vertical. The effect of aerosol burden is examined by varying CCN concentration over which cloud droplets are activated at 1% supersaturation in the Morrison two-moment bulk microphysics scheme. Radiative heating is held prescribed and noninteractive at a constant rate in the atmospheric column, summing vertically to 122 W m^{-2} , to completely isolate aerosol-microphysical effects on convection and precipitation. Analyses are conducted only on the equilibrated part of the simulations.

Mean precipitation shows a small monotonic reduction with weaker convection as a function of increasing CCN concentration. Analysis of the vertically integrated moist static energy budget indicates that the decrease in convection and rainfall is attributed to the reduction in surface enthalpy fluxes, as the top-heavy divergent circulation maintains the same vertical shape and the gross moist stability remains constant.

Increasing CCN concentration changes the domain-mean cloud liquid and ice water mixing ratio and redistributes heat and moisture in the atmospheric column. Atmospheric temperature cools above 850 hPa and is associated with less diabatic heating, and anomalously descending W . In the lower levels, warmer and drier air reduces evaporative cooling and decreases the air-sea temperature disequilibrium, which, in turn, reduces the latent and sensible heat flux and decreases surface precipitation.

It is important to note that only when changes in the column temperature and moisture, as caused by increasing CCN concentration, are coupled to the large-scale dynamics (so that cloud diabatic heating feeds back on the large-scale adiabatic motion and vice versa) are changes in surface precipitation manifested, as sensitivity tests show. Our results indicate that weaker convection and reduction in mean rainfall are primarily caused by the feedback from convective coupling with the large-scale dynamics, and not by the effect of enhanced aerosols on cloud droplets and microphysical properties. Nonetheless, the microphysical picture remains consistent with the macrophysical counterpart.

Enhanced CCN concentration results in an increase in the number concentration of cloud droplets, shifting the droplet size toward smaller radii for a given liquid water content. Particle collision becomes less efficient and suppresses rain production, which increases cloud liquid and ice water mixing ratio in the troposphere.

A few caveats remain to be further explored, particularly regarding the treatment of supersaturation in the bulk microphysics schemes. Schemes that use saturation adjustment, like the one used here, were shown to produce stronger cloud buoyancy than schemes with predicted supersaturation (Grabowski and Morrison 2017). Although the two-moment bulk schemes suffer from several limitations in the treatment of aerosol–cloud interaction (e.g., Khain et al. 2015; Fan et al. 2016), we do not expect spectral bin microphysics schemes that explicitly predict supersaturation, to give a qualitatively different answer. This is because at tropical quasi equilibrium, convection is controlled by the interaction between cloud buoyancy and the large-scale vertical motion though gravity waves adjustment. Therefore, when allowing convective adjustment feedback microphysical effects from droplet size and concentration become irrelevant as we show in a series of sensitivity tests.

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