Effect of Surface Fluxes versus Radiative Cooling on Tropical Deep Convection.

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ABSTRACT

The effects of turbulent surface fluxes and radiative cooling on tropical deep convection are compared in a series of idealized cloud-system resolving simulations with parameterized large scale dynamics. Two methods of parameterizing the large scale dynamics are used; the Weak Temperature Gradient (WTG) approximation and the Damped Gravity Wave (DGW) method. Both surface fluxes and radiative cooling are specified, with radiative cooling taken constant in the vertical in the troposphere. All simulations are run to statistical equilibrium.

In the precipitating equilibria, which result from sufficiently moist initial conditions, an increment in surface fluxes produces more precipitation than equal increment of column-integrated radiative cooling. This is straightforwardly understood in terms of the column-integrated moist static energy budget with constant normalized gross moist stability. Under both large-scale parameterizations, the gross moist stability does in fact remain close to constant over a wide range of forcings, and the small variations which occur are similar for equal increments of surface flux and radiative heating.

With completely dry initial conditions, the WTG simulations exhibit hysteresis, maintaining a dry state with no precipitation for a wide range of net energy inputs to the atmospheric column. The same boundary conditions and forcings admit a rainy state also (for moist initial conditions), and thus multiple equilibria exist under WTG. When the net forcing (surface fluxes minus radiative cooling) is increased enough that simulations which begin dry eventually develop precipitation, the dry state persists longer after initialization when the surface fluxes are increased than when radiative cooling is
decreased. The DGW method, however, shows no multiple equilibria in any of the simulations.

1. Introduction

Surface turbulent heat fluxes and electromagnetic radiation are the most important sources of moist static energy (or moist entropy) to the atmosphere. In the idealized state of radiative-convective equilibrium (RCE), the source due to surface fluxes must balance the sink due to radiative cooling. In this state, the surface evaporation and precipitation also balance, and there is no large-scale circulation. In a more realistic situation in which there is a large-scale circulation, the strength of that circulation’s horizontally divergent component can be viewed as proportional, in a column-integrated sense, to the net moist static energy source (surface fluxes minus column-integrated radiative cooling), with the proportionality factor being known as the gross moist stability.

There is no accepted theory which satisfactorily predicts the gross moist stability. It is itself a function of the large-scale circulation, and may be dynamically varying. If we could be certain that it would remain constant, however, then not only would the divergent circulation (i.e., the large-scale vertical motion) be predictable as a function of the surface fluxes and radiative cooling, but surface fluxes and radiative cooling would influence that circulation in the same way. All that would matter would be the difference between the two, the column-integrated net moist static energy forcing. This study investigates, in an idealized setting, whether this is the case. We ask whether surface fluxes and radiation influence the circulation differently. We might expect that they would, given that surface fluxes act at the surface while radiation acts throughout the column. Such a difference would necessarily be expressed (at least in the time mean) as
a difference in the gross moist stability between two situations in which the net moist static energy source is the same, but its partitioning between surface fluxes and radiation is different.

We study this problem using a Cloud Resolving Model (CRM). CRMs have proven to be very powerful tools for studying deep moist convection. One set of useful studies involves simulations of RCE (e.g., Emanuel 2007, Robe and Emanuel 2001, Tompkins and Craig 1998a, Bretherton et al. 2005, Muller and Held 2012, Popke et al. 2012, Wing and Emanuel 2014). While RCE has provided many useful insights, it entirely neglects the influences of the large scale circulation. Another approach is to parameterize the large scale circulation (e.g., Sobel and Bretherton 2000; Mapes 2004; Bergman and Sardeshmukh 2004; Raymond and Zeng 2005; Kuang 2011; Romps 2012; Wang and Sobel 2011; Anber et al. 2014; Edman and Romps 2014) as a function of variables resolved within a small domain. This approach is computationally inexpensive (compared to using domains large enough to resolve the large scales present on the real earth), and provides a two-way interaction between cumulus convection and large scale dynamics.

One method to parameterize the large scale dynamics is called the Weak Temperature Gradient (WTG) approximation (Sobel and Bretherton 2000). As the name suggests, this method relies explicitly on the smallness of the temperature gradient in the tropics, which is a consequence of the small Coriolis parameter there. Any temperature, or density, anomaly in the free troposphere generated by diabatic processes is rapidly wiped out by means of gravity wave adjustment to restore the temperature profile to that
of the adjacent regions. Hence the dominant balance is between diabatic heating and adiabatic cooling, and the tropospheric temperature is constrained to remain close to a target profile which is interpreted as that of surrounding regions.

While WTG captures the net result of the gravitational adjustment, it does not simulate the gravity waves themselves. Another method of representing the large scale dynamics in CRMs represents those dynamics as resulting explicitly from such waves, with a single wavenumber, interacting with the simulated convection. This method was introduced by Kuang (2008) and Blossey et al. (2009) and is called the Damped Gravity Wave (DGW) method.

Both methods have been shown to produce results qualitatively similar to observations in some settings; for example, Wang et al. (2013) compared the two methods with observations produced during the TOGA-COARE field experiment. Utilizing both of the above two methods, as we do here, allows us to explore a variety of mechanisms and parameters affecting the interaction between deep convection and large scale dynamics, among which are the surface turbulent fluxes and radiative cooling.

In numerical experiments using the WTG method, Sobel et al. (2007) and Sessions et al. (2010) found that the statistically steady solution is not unique for some forcings: the final solutions can be almost entirely dry, with zero precipitation, or rainy, depending on the initial moisture content. We have interpreted this behavior as relevant to the phenomenon of “self-aggregation” in large-domain RCE simulations (Bretherton et al. 2005; Muller and Held; Wing and Emanuel 2014), with the two states corresponding to dry and rainy regions within the large domain. Tobin et al. (2012) find evidence of this behavior in observations. In the present study, we perform sets of simulations with
different initial conditions to look for multiple equilibria, and to determine whether their existence or persistence is influenced differently by surface fluxes and radiation.

This paper is organized as follows: in section 2 we describe the model and the experiment setup. In section 3 we show results. We highlight some implications of our results and conclude in section 4.

2. Model configuration and experimental setup:

2.1. Model configuration:

We use the Weather Research and Forecast (WRF) model version 3.3, in three spatial dimensions, with doubly periodic lateral boundary conditions. The experiments are conducted with Coriolis parameter $f = 0$. The domain size is $192 \times 192 \text{ km}^2$, with a horizontal grid spacing of 2 km. There are 50 vertical levels in the domain, extending to 22 km high, with 10 levels in the lowest 1 km. Gravity waves propagating vertically are absorbed in the top 5 km to prevent unphysical wave reflection off the top boundary using the implicit damping vertical velocity scheme (Klemp et al. 2008). The 2-dimensional Smagorinsky first-order closure scheme is used to parameterize the horizontal transports by sub-grid scale eddies. The Yonsei University (YSU) first order closure scheme is used to parameterize boundary layer turbulence and vertical subgrid scale eddy diffusion (Hong and Pan 1996; Noh et al. 2003; Hong et al. 2006). The microphysics scheme is the Purdue-Lin bulk scheme (Lin et al. 1983; Rutledge and Hobbs 1984; Chen and Sun 2002) that has six species: water vapor, cloud water, cloud ice, rain, snow, and graupel.
We first perform an RCE experiment at fixed sea surface temperature of 28 °C until equilibrium is reached at about 60 days. Results from this experiment are averaged over the last 10 days to obtain statistically equilibrated temperature and moisture profiles. Figure 1 shows the resulting vertical profiles of (a) temperature and (b) moisture. These profiles are then used to initialize other runs with parameterized large scale circulations, and the temperature profile is used as the target profile against which perturbations are computed in both the WTG and DGW methods. We will call the RCE moisture profile the non-zero moisture profile, or wet conditions, to distinguish it from other moisture profiles (zero, in particular, or dry conditions) used in this paper.

2.2. Parameterized large scale circulation:

The large scale vertical velocity is dynamically determined using either the WTG or the DGW method. In the relaxed form of WTG used in CRM simulation (Raymond and Zeng 2005; Wang and Sobel 2010; Wang et al. 2013; Anber et al. 2014) the vertical velocity $W$ is obtained by:

$$W(z) = \begin{cases} 
\frac{1}{\tau} \frac{\theta - \theta_0}{\partial \theta/\partial z} & ; z \geq h \\
\frac{z}{h} W(h) & ; z < h 
\end{cases}$$

where $\theta$ is the domain mean potential temperature, $\theta_0$ is the reference temperature (from RCE run), $h$ is the height of the boundary layer determined internally by the boundary
layer scheme, and $\tau$ is the relaxation time scale, and can be thought of as the time scale
over which gravity waves propagate out of the domain, taken here 3 hours.

In DGW method (Kuang 2008; Blossey et al 2009; Romps 2012a, 2012b; Wang et al. 2013) the large scale vertical velocity is obtained by solving the elliptic partial
differential equation:

$$\frac{\partial}{\partial p} (\varepsilon \frac{\partial \omega}{\partial p}) = \frac{k^2 R_d}{p} (T_v - T_v^0) \quad \ldots(2)$$

where $p$ the pressure, $\omega$ is the pressure vertical velocity, $R_d$ is the dry gas constant, $T_v$ is
the domain mean virtual temperature, $T_v^0$ is the target virtual temperature (from RCE), $\varepsilon$
is the momentum damping, in general a function of pressure but here taken constant at
1 day$^{-1}$, and $k$ is the wavenumber taken $1.6 \times 10^{-5}$ m$^{-1}$.

The boundary conditions used for solving (2) are:

$$\omega(0) = \omega(100 \ hpa) = 0$$

Once the vertical velocity obtained from (1) or (2), it is used to vertically advect domain
mean temperature and moisture at each time step. Horizontal moisture advection is not
represented.

The free parameters used here are chosen to give a reasonable comparison between the
general characteristics of the two methods, and to produce a close, but not exact,
precipitation magnitude in the control runs.
2.3. Experiment design

All simulations are conducted with prescribed surface fluxes and radiative cooling and no mean wind. Radiative cooling is set to a constant rate in the troposphere, while the stratospheric temperature is relaxed towards 200 K over 5 days as in Wang and Sobel (2011) and Anber et al. (2014).

The control runs have surface fluxes of 205 Wm$^{-2}$; latent heat flux (LH) of 186 Wm$^{-2}$ and sensible heat flux (SH) of 19 Wm$^{-2}$, (the ratio of the two corresponding to Bowen ratio of 0.1) and vertically integrated radiativecooling of 145 Wm$^{-2}$, corresponding to a radiative heating rate of -1.5 K/day in the troposphere in both the WTG and DGW experiments.

(Note that we use both the terms “radiative cooling” and “radiative heating” although one is simply the negative of the other. The radiative heating is always negative in our simulations and thus is most simply described as “radiative cooling”, but when we compare different simulations a positive change – an increase in radiative heating – is directly compared to a positive change in surface turbulent heat fluxes, and thus when such changes are described it is simpler to describe such changes in terms of radiative heating.)

We perform two sets of experiments with parameterized large scale dynamics: one in which surface fluxes are varied by increasing or decreasing their prescribed magnitude by 20 Wm$^{-2}$ from the control run while holding radiative cooling fixed at 145 Wm$^{-2}$, and the other in which the prescribed radiative cooling is varied in increments of 20 Wm$^{-2}$ while holding surface fluxes fixed. Perturbations in $Q_R$ are performed by
varying radiative cooling rate while holding in uniform in the vertical. Table 1 summarizes the control parameters of the numerical experiments.

Another two sets of simulations (with two methods) are performed which are identical except that they are initialized with zero moisture profile (or “dry conditions”). All mean quantities are plotted as a function of the net energy input (NEI) to the atmospheric column excluding the contribution from circulation. Thus, NEI is the sum of surface fluxes (SF) and vertically integrated radiative heating ($Q_R$): $\text{NEI} = \text{SF} + (Q_R)$.

3. Results:

3.1. Precipitation and Normalized Gross Moist Stability

a. Mean precipitation:

a.1. Non-zero initial moisture conditions:

Figure 2 shows the domain and time mean precipitation as a function of the net energy input (NEI) using the non-zero moisture profile as initial condition with (a) WTG and (b) DGW. At zero NEI, in one set of (red) experiments the radiative cooling rate is reduced from that in the control to balance surface fluxes (205 W m$^{-2}$); while in the other (blue) surface fluxes is reduced to balance radiative cooling (145 W m$^{-2}$). The former gives more precipitation in both WTG and DGW experiments.

In all these experiments, the precipitation rate varies linearly over a broad range of NEI values. The precipitation rate produced for a given increment of surface fluxes exceeds that produced for the same increment of vertically integrated radiative heating (equivalently, the opposite increment in radiative cooling). For example, increasing surface fluxes by 40 W m$^{-2}$ from the control run (i.e. at 100 W m$^{-2}$ or surface fluxes
exceeds radiative cooling by 100 Wm$^{-2}$) there is more precipitation (blue curve) than if we decrease radiative cooling by 40 Wm$^{-2}$ (red curve).

It is straightforward to understand this difference in the slopes of precipitation responses from the point of view of the column-integrated moist static energy budget. We use the steady state diagnostic equation for precipitation as in, e.g., Sobel (2007), Wang and Sobel (2011), or Raymond et al. (2009):

$$ P = \frac{1}{M} (L + H + \langle Q_R \rangle) - \langle Q_R \rangle - H \quad ...(3) $$

Where $\langle . \rangle = \int_{P_0}^{P_T} \frac{dp}{g}$ is the mass weighted vertical integral from the bottom to the top of the domain. $P$, $L$, $H$, $SF$ and $Q_R$ are precipitation, latent heat flux, sensible heat flux, surface fluxes (sum of latent and sensible heat flux), and radiative heating.

We define $M = \left\langle W \frac{\partial \tilde{h}}{\partial z} \right\rangle / \left\langle W \frac{\partial \tilde{s}}{\partial z} \right\rangle$ as the normalized gross moist stability, which represents the export of moist static energy by the large-scale circulation per unit of dry static energy export (e.g., Neelin and Held 1987; Sobel 2007; Raymond et al. 2009; Wang and Sobel 2011; Anber et al 2014). Here $h$ is the moist static energy (sum of the thermal, potential and latent energy), $s$ is the dry static energy (thermal and potential energy), and the overbar is the domain mean and time mean. The second and third terms (combined) on the right hand side of (3) represent the precipitation that would occur in radiative convective equilibrium. The first term accounts for the contribution by the large scale circulation, which arises from the discrepancy between surface fluxes and vertically integrated radiative cooling. Therefore,
\( \langle Q_R \rangle \) contributes to \( P \) in two ways with opposite signs; to the dynamic part (the first term on the right hand side of (3)), similar to the contribution from surface fluxes, and to the RCE precipitation (the second term on the RHS of (3)) in an opposite sense. Surface fluxes, on the other hand, contribute only positively.

Figure 3 shows the normalized gross moist stability (\( M \)) as a function of \( \text{NEI} > 0 \) for cases initialized with non-zero initial moisture conditions from (a) WTG and (b) DGW experiments. \( M \) is a positive number less than 1 and remains close to constant under each forcing method, though the values under DGW are consistently smaller than those under WTG. The smallness of the variations in \( M \) is a nontrivial result; we know no \textit{a priori} reason why \( M \) could not vary more widely. Even the variations which do occur as a function of \( \text{NEI} \) are similar for equal increments of surface flux or radiative heating, over most of the range, particularly in DGW. The most marked differences occur at \( \text{NEI} = 20 \text{ Wm}^{-2} \) under WTG, the value closest to RCE.

At \( \text{NEI} = 0 \) the large scale vertical velocity vanishes and \( M \) is undefined; however \( M \) in that case is not needed to compute \( P \). Equation (3) is derived by eliminating the vertical advection term between the moist and dry static energy equations, but \( \text{NEI} = 0 \) corresponds to RCE, in which the vertical advection vanishes. In that case, the precipitation is simply \( P = \langle -Q_R \rangle - H \).

When there is a large-scale circulation such that (3) is valid, we can see that if \( M \) and the surface fluxes are held fixed, the change in precipitation per change in radiative cooling is:
\[ \frac{\partial P}{\partial \langle Q_R \rangle} = \frac{1}{M} - 1 \quad \text{...(4)}. \]

As discussed above, Figure 3 shows that constancy of M is a good approximation for all the numerical experiments.

On the other hand, the change in precipitation due to an increment in surface fluxes (holding radiative cooling and M fixed) scales as:

\[ \frac{\partial P}{\partial SF} = \frac{1}{M} \quad \text{...(5)}. \]

Equations (4) and (5) show that a change in precipitation due to an increment in surface fluxes will exceed that due to an increment in radiative cooling. The difference of unity, nondimensionally, means that for finite and equal increments of either surface fluxes or radiative heating, the excess precipitation due to surface fluxes is equal to the increment in forcing itself.

Given a positive M, equation (5) states that increasing surface fluxes always increases precipitation, but precipitation responses to changes in \( \langle Q_R \rangle \) can be either negative or positive in principle, depending whether M is greater or less than 1. For a small M (\( M \ll 1 \)), the difference is small. In our experiments, where M is sufficiently large (~0.4, as shown below) that the difference is not negligible, surface fluxes have a significantly greater influence on precipitation than does radiative cooling.

The difference we see in FIG.2 is what we expect for constant M, and might have been considered a null hypothesis. It indicates no fundamental difference in how surface fluxes and radiative cooling influence the large scale circulation in these simulations. We
might have expected that forcing in the interior of the troposphere by radiative cooling
might induce differences in the thermodynamic profiles and the vertical motion profiles
relative to forcing at the surface by turbulent fluxes, resulting in different values of $M$.
However, the simulations here are well explained by the simplest vertically integrated
theory, in which $M$ remains constant. The simulations are needed, in this sense, only to
provide the (single) value of $M$.

Quantitatively speaking, the slope of the rainfall in Figure 2.a (WTG, and
similarly for DGW) for increasing the surface fluxes, $\frac{\partial P}{\partial SF}$ is about 2.3, which
corresponds to $M = 0.44$ using equation (5). Similar magnitudes are obtained for $\frac{\partial P}{\partial (Q_R)}$.
For DGW, however, $\frac{\partial P}{\partial SF}$ is about 3.1, and corresponds to $M = 0.32$. Both values of $M$ are
close to those obtained in the control runs with $NEI = 60$ Wm$^{-2}$.
As in Figure 3, the right hand side of (3) with constant $M$ gives a very good estimate of
the changes in precipitation resulting from both types of forcings.

2. Zero initial moisture conditions:
Figure 4 is as analogous to Figure 2 but now we initialize the simulations with a
zero moisture profile while keeping everything else the same as above (including the
reference temperature profile). For WTG (Figure 4.a) the system now exhibits multiple
equilibria, staying in the dry state over a wide range of NEI (Sobel et al. 2007, Sessions et
al. 2010). Precipitation does not occur for these dry initial conditions for NEI below a
threshold value. Above this value, we obtain mean precipitation values identical to those found with non-zero initial moisture profile for the same forcings. However, the transition to a precipitating state (that is, to the apparent inability of the dry state to be sustained) occurs at 60 Wm$^{-2}$ when radiative cooling is varied, but 70 Wm$^{-2}$ when surface fluxes are varied. In other words, starting from dry conditions the system requires less energy from radiation than from surface fluxes in order for precipitation to occur.

Under the DGW method (Figure 4.b), however, the system produces the exact same precipitation rate as was produced with the non-zero initial moisture profile as an initial condition. This suggests that the DGW method does not allow multiple equilibria in the parameter range we have explored.

b. Precipitation Time Series:

In this section we explore whether there are any differences between the influences of surface fluxes and radiative cooling in the simulations starting with dry conditions other than those evident in the statistical equilibria which are eventually reached. Specifically, in the simulations where precipitation eventually does occur – but near the threshold value of NEI below which the dry state can be sustained – we ask whether the time interval between the initial time and the time at which precipitation first occurs may be different.

Figure 5 shows domain mean precipitation time series under WTG for the dry initial conditions with NEI of (a) 70 (b) 80, and (c) 90 Wm$^{-2}$, all of which are near the transition from existence to non-existence of the dry solution (Fig. 4a). Two points are worth noticing here. First, the transition from the dry to precipitating state happens in a
more dramatic fashion, overshooting the statistical equilibrium value, when radiative
cooling is perturbed. In contrast, perturbing surface fluxes leads to a much smoother
transition to precipitation onset. Second, the precipitation lag due to perturbing surface
fluxes vs. radiative cooling is apparent, not only in the time mean picture, but also in the
time series, as the time interval between initialization and precipitation onset is longer for
a surface flux increase than an equivalent radiative heating increase. This lag decreases as
the NEI increases (case of 100 Wm$^{-2}$ shows almost no lag, hence not shown).

In Figure 5.a, case of NEI = 70 Wm$^{-2}$ in which precipitation occurs only from a radiative
cooling reduction, the onset of precipitation is delayed by about 35 days, after which it
takes less than 3 days to reach equilibrium.

The delay in precipitation onset is reduced when NEI is increased. For NEI = 80 Wm$^{-2}$
(Figure 5.b), the delay is about 20 days for the radiative cooling perturbation and about
27 days for the surface flux perturbation. For the case of NEI = 90 Wm$^{-2}$ (Figure 5.c),
delays are about 16 days and 18 days for radiative cooling and surface flux perturbations,
respectively.

The precipitation time series for the case NEI = 40 Wm$^{-2}$ in the DGW experiment
with a dry initial moisture profile is shown in Figure 6. Unlike the WTG experiment, the
time the system takes to begin precipitating is indistinguishable for radiative cooling and
surface flux perturbations (the same is true for other NEI cases, not shown). This might
be due to the vertical structure of the large scale vertical velocity in WTG as we will
discuss in the next section. When starting from dry conditions, precipitation is delayed
for only 5 days compared to non-zero initial moisture conditions, and this lag is not
dependent on the NEI.
3.2. Large Scale Vertical Velocity:

We now focus on the vertical profiles of large scale vertical velocity, $W$. We first examine these for the experiments with moist initial conditions. Time mean vertical profiles of $W$ in the precipitating equilibrium under WTG and DGW and their corresponding maximum values are shown in Figures 7 and 8, respectively, for experiments in which surface fluxes are perturbed. Time mean $W$ from the experiments $\text{NEI} = 0$ is not shown because it is zero by design. As expected, the large scale vertical velocity is more top heavy under WTG than DGW (Romps 2012a and b; Wang et al, 2013). The profiles’ peak values are almost identical for the same increment in radiative cooling despite different precipitation magnitudes.

All these experiments produce moist states with non-zero precipitation (Figure 2). (Even moderate rates of subsidence need not, generally, be associated with completely dry states; if the condensation heating is nonzero but smaller than the radiative cooling, descent will still occur.) When the NEI is negative, vertical motion is downward and precipitation falls below the RCE magnitude. The opposite happens when the NEI is positive.

The vertical structure of $W$ also offers some clues to the precipitation onset in the experiments with dry initial conditions. Figure 9 shows the time evolution of the vertical structure of WTG vertical velocity for the experiments initialized with dry conditions due to perturbations in (a) surface fluxes and (b) radiative cooling, for the case of $\text{NEI} = 90 \text{Wm}^{-2}$ (other cases are similar and not shown). Consistent with the time evolution of precipitation (Figure 5.c), ascending vertical motion is delayed until about the same time
the system precipitates. When radiative cooling is perturbed, ascending motion is strong in the boundary layer at the transition time from the dry to wet state (day 16 in Figure 9.b). This might explain the dramatic transition in the \( \langle Q_r \rangle \) perturbation experiments (Figure 5).

One can think of radiative cooling as inhibiting convection through subsidence, which in turn induces drying which keeps relative humidity low. In an atmosphere with low relative humidity, convection is inhibited when entrainment mixes dry air into saturated updrafts, reducing buoyancy. We hypothesize that reducing the radiative cooling is quicker to eliminate the dry equilibrium than increasing surface fluxes because any change in radiative cooling has an impact on humidity (and thus on the environment for deep convection) throughout the troposphere, whereas the effects of surface fluxes can be trapped in the planetary boundary layer if the troposphere is dry.

3.3. Thermodynamic Variables:

a. Temperature anomalies:

In order to induce ascending motion at any given level, mean temperature anomalies, relative to the reference temperature profile, must be positive under WTG by (1). This is also true, in a more approximate and vertically nonlocal sense, under DGW. Figure 10 shows the time evolution of the vertical structure of potential temperature anomalies (from the RCE target temperature profile) for the case of NEI = 80 W m\(^{-2}\) (when surface fluxes are perturbed by 80 W m\(^{-2}\) from the control run) initialized from the dry conditions using either (a) WTG or (b) DGW method. For WTG, dry conditions are maintained by colder free tropospheric temperature (negative anomalies). A transition to
a wet state requires positive free tropospheric temperature anomalies to generate ascending motion. Such anomalies develop early in all our simulations under the DGW method. Under WTG (Figure 10a), after the initialization but before precipitation onset, warm anomalies build up in the lowest 1 km while cold anomalies are maintained above. It appears that the buildup of low level warm anomalies contributes to greater conditional instability which is eventually realized in a rapid onset of convection.

Unlike under WTG, the smooth vertical structure of the large scale vertical velocity under the DGW method makes it more efficient at communicating temperature anomalies between the boundary layer and the free troposphere, since pressure and temperature adjust non-locally through the hydrostatic balance. This may explain why DGW experiments appear to reach a unique statistically steady state.

b. Relative humidity (RH):

Figure 11 shows the time evolution of the vertical structure of relative humidity for cases initialized with dry conditions due to perturbations in (a) surface fluxes and (b) radiative cooling, with NEI = 90 Wm\(^{-2}\) (other cases are similar and not shown) in the WTG experiments. The most striking feature is that for the radiative cooling perturbations, there is a build-up of relative humidity exceeding 90% at approximately 2 km altitude, deepening the moist layer above the surface, about 5 days prior to the onset of precipitation (denoted by the white dashed vertical line). This prior RH build-up does not occur when surface fluxes are perturbed; in Figure 11a, the RH just above the boundary layer increases substantially only after precipitation has begun, rather than before as in Figure 11b. Further, the free troposphere as a whole takes a couple of days after precipitation onset to moisten in Figure 11a whereas it moistens very quickly after
precipitation onset in Figure 11b; this suggests a more active role for free tropospheric
humidity in causing (or at least allowing) the onset of precipitating convection for the
radiative heating increase, whereas the tropospheric humidification appears to be more
clearly a consequence of the convection for the surface flux increase Though a subtle
difference, this supports the hypothesis that radiative cooling reductions are quicker to
eliminate the dry equilibrium than surface flux increases because they have different
effects on free-tropospheric humidity in the dry equilibrium.

Under the DGW method, on the other hand, there is no such RH build-up (not
shown) when increments in either radiative heating or surface fluxes are introduced.

3.4. Sensitivity experiments

We have done simulations in which the initial moisture profile is neither equal to
the RCE profile nor zero, but different fractions of the RCE initial moisture profile. For
moisture profiles greater than approximately 50% of the RCE profile, only the rainy
equilibrium state is reached. For drier profiles, the dry equilibrium state can be reached,
though over a narrower range of NEI than when the completely dry initial conditions are
used.

We also have done simulations with interactive surface fluxes with specified sea
surface temperature. We obtain similar results to those with fixed surface fluxes if we
compare simulations in which the actual values of the fluxes are similar - including the
existence or non-existence of multiple equilibria - as long as the NEI remains in the
interval (-20, 60) Wm$^{-2}$. 
Finally, we performed ensemble simulations similar to those described here for the zero moisture initial conditions but with perturbed (positive and negative) temperature and wind initial conditions. Variations in precipitation onset near the transition from dry to wet state are less than a day for the experiments with perturbed radiative cooling, but identical in all other experiments. This indicates that the differences in response are due to differences in forcing and not due to random variability.

4. Conclusions

Cloud Resolving Model simulations have been conducted with parameterized large scale circulation to contrast the effects of surface fluxes and radiative cooling on deep tropical convection. Two different parameterizations of large-scale circulation, the weak temperature gradient (WTG) and damped gravity wave (DGW) methods, are used. In the precipitating equilibrium state, a given change in surface fluxes induces a greater change in precipitation in our simulations than does an equal change in radiative heating. This difference is a straightforward consequence of the column-integrated moist static energy budget with a constant normalized gross moist stability. The surface flux and radiative cooling increments result in equal changes in the divergent circulation. The precipitation change, however, is a consequence of both that divergent circulation change and the change which would occur in its absence – that is, in radiative-convective equilibrium (RCE). In RCE, surface fluxes and radiative heating have opposite effects on precipitation; the overall precipitation change in our simulations is thus the sum of these
opposite contributions from the RCE component and equal contributions from the induced circulations.

Aside from the differences in precipitation that result straightforwardly from their different roles in RCE, however, equal increments of surface fluxes and radiative heating influence our simulations identically for all practical purposes. The large-scale vertical motion and moist static energy export changes induced by equal increments of the two forcings are essentially identical. We might have expected, on the contrary, that the two forcings would induce different responses in such a way that the normalized gross moist stabilities would be different, allowing differences in large-scale vertical motion.

This does not occur; the gross moist stability, under both WTG and WTG, remains approximately constant under each method (though it is modestly smaller under DGW than WTG due to the less top-heavy vertical motion profiles). Even the small variations which do occur are similar (over most of the range studied) for equal surface flux and radiation perturbations. If the normalized gross moist stability were precisely constant, the precipitation rate in all the experiments (under a given large-scale parameterization) could be predicted accurately from the moist static energy budget (3) after doing a single simulation to determine the gross moist stability, since the surface fluxes and radiative cooling are both specified in these simulations. While this is a somewhat unrealistically constrained situation compared to the real one in which surface fluxes and radiation are interactive, it is nonetheless interesting that the one degree of freedom our simulations do have in the column integrated moist static energy budget – the normalized gross moist stability – is exercised almost not at all.
A set of simulations was also conducted in which the model was initialized with zero moisture, to determine whether multiple equilibria exist and whether surface fluxes and radiation affect their influence differently. Under WTG, a dry non-precipitating equilibrium can be maintained over a wide range of NEI. To make a transition to a wet state, the system needs a smaller increase in radiative heating than surface fluxes. We interpret this as resulting from the distribution of radiative cooling through the whole atmospheric column, such that reducing it reduces subsidence and increases humidity above the planetary boundary layer. Consistent with this, a deepening of the moist layer adjacent to the surface occurs prior to the onset of precipitation in the simulation forced by a radiation change which does not occur in the simulation forced by an equal surface flux change (in which the onset of precipitation occurs later).

In the DGW method, only a precipitating equilibrium state is found. This may be understood (in the sense of proximate causes) in terms of the warm temperature anomalies in the free troposphere produced by this method that cause ascending large scale vertical motion.

Studies of self-aggregation of convection in large-domain-CRMs (Bretherton et al. 2005; Muller and Held 2012; Jeevanjee and Romps 2013; Wing and Emanuel 2014; Emanuel et al. 2014) show that interactive radiation is essential to the occurrence of self-aggregation. The multiple equilibria occurring in single-column or small-domain CRM simulations under WTG (e.g., Sobel et al 2007, Sessions et al. 2010) have been interpreted as a manifestation of the same phenomenon, yet interactive radiation is not required for its occurrence in our WTG simulations (or in those of Sessions et al. 2010). Herman and Raymond (2014) show that the occurrence of multiple equilibria in WTG
simulations without interactive radiation is sensitive to the choice of the level used for the boundary layer top, a free parameter in the method, and that multiple equilibria do not occur in a new spectral WTG method which – similarly to the DGW method – is nonlocal in the vertical and does not require a special treatment of the boundary layer. Their results and ours appear broadly consistent, and suggest that the occurrence of multiple equilibria without interactive radiation under standard WTG method may be an artifact of the method’s locality in the vertical or (relatedly) its somewhat ad hoc treatment of the boundary layer. Which (if any) of these methods produces multiple equilibria in the presence of interactive radiation in CRMs, and whether the dynamics of those equilibria are faithful to the dynamics of self-aggregation seen in large-domain RCE simulations, remains as a question for future work.
Acknowledgement

This work was supported by NSF grant AGS-10088 47. We would like to acknowledge high-performance computing support from Yellowstone (ark:/85065/d7wd3xhc) provided by NCAR's Computational and Information Systems Laboratory, sponsored by the National Science Foundation.
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FIG.4. Same as Figure 2 but using zero moisture initial conditions.

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TABLE 1. Control parameters of net radiation, surface fluxes and their components of sensible and latent heat flux (chosen to keep the Bowen ratio close to 0.1). The sum of surface fluxes and radiative cooling is the Net Energy Input (NEI). All the numbers are in Wm$^{-2}$. Bold marks the control run parameters (see text for more details on the experiment setup). NEI of 70 and 90 Wm$^{-2}$ are for cases initialized with dry conditions using WTG method.

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<th>NEI</th>
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<th>Sensible Heat</th>
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