Explicitly simulated tropical convection over idealized warm pools

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\[1\] Two-dimensional cloud-system-resolving simulations of convection over idealized warm pools are conducted to examine the relationship between the spatial patterns of tropical convection and the sea surface temperature (SST) distributions. Results show that the most active convection resides near the edge of warm pools, with a local minimum around the warmest center. This finding might provide an interpretation for the observations that peak convection is commonly located several degrees of latitude off the maximum SST over some tropical oceans. Factors potentially affecting the convective patterns are explored through sensitivity experiments. It is found that convection expands significantly and rainfall peaks are further displaced several hundred more kilometers away from the warmest water when the radiative cooling/heating is applied homogeneously across the domain. Conversely, when the wind-induced surface flux variability is excluded, convective activity is confined within a much narrower area of high SSTs, but the overall spatial pattern is largely retained. Moreover, the surface friction exerts profound effects on the simulated convection and attendant large-scale flow and is mostly responsible for the dual-maximum precipitation and two-cell circulation structure in the horizontal. These results suggest that as well as the temperature/pressure gradients resulting from the non-uniform SSTs, other processes must be taken into consideration in the interpretation of observed tropical convection and circulation.


1. Introduction

\[2\] Tropical convection is of central importance to both tropical and global atmospheric circulations. Many critical aspects of its large-scale organizations, however, are still not fully understood. Among these issues is how the spatial and temporal distribution and variability of deep precipitating convection are modulated by the sea surface temperature (SST) over the tropical oceans.

\[3\] Empirical SST-convection relationships have been examined by the pointwise analysis in a number of observational studies [e.g., Graham and Barnett, 1987; Zhang, 1993; Arking and Ziskin, 1994; Fu et al., 1994; Lau et al., 1997; Halpern and Hung, 2001]. These previous studies consistently reveal that strong convection tends to reside over warm sea surfaces; especially, a dramatic change from nonconvective to convective regimes appears around a SST threshold of roughly 26.5\textdegree{}–27.5\textdegree{}C. The observed preference of convective activity for high SST is physically interpretable in terms of the large-scale surface moisture convergence induced by SST gradients [Lindzen and Nigam, 1987], as well as the influence of local SSTs on the atmospheric boundary-layer.

\[4\] However, observational analysis demonstrates that considerable variation exists in the relationship between tropical deep convection and the underlying SST distribution. For example, deep convection tends to maximize at about 29.5\textdegree{}–30\textdegree{}C and then decrease with further increasing SST [Zhang, 1993; Waliser and Graham, 1993]. Also, situations with no and weak deep convection can be observed at high SST, and furthermore, the highest SST is often not colocated with the maximum convection [e.g., Ramage, 1974; Sadler, 1975; Hastenrath and Lamb, 1977; Lietzke et al., 2001]. These complicated convection-SST relationships possibly arise from the interplay of different physical mechanisms, including both SST distribution and the large-scale atmospheric circulation [Fu et al., 1994; Lau et al., 1997], and they are poorly understood.

\[5\] The relationship between convective activity and SST has also been a topic of a few numerical modeling studies. For instance, using cloud-resolving models (CRMs), Lau et al. [1993] and Tompkins and Craig [1999] investigated the dependence of tropical convection on local SSTs with and without large-scale forcing, respectively. Their main conclusion is that convection is largely insensitive to the underlying uniform SST changes. Raymond [1994] performed simulations of a two-dimensional large-scale tropical circulation driven by SST gradients using parameterized convection. A similar mock-Walker circulation problem was...
considered by Grabowski et al. [2000, hereafter, GYM] using the cloud-resolving modeling approach. These studies showed that the maximum convection coincides with the warmest water, consistent with the hypothesis by Lindzen and Nigam [1987]. However, we will show that these modeling results could be an artifact of either the absence of some critical physical processes or the use of traditional cumulus parameterizations.

In this study, we revisit the tropical deep convection and resultant large-scale circulation in response to a prescribed SST distribution in a context of cloud-system-resolving modeling. The primary objective is to identify the spatial patterns of convective activity solely under the influence of underlying sea surface conditions, namely to isolate the role of SSTs in regulating convection from other factors. We also quantify the role of cloud-radiation interaction, surface wind variability, surface friction, and SST anomaly patterns in regulating the simulated tropical convection through sensitivity experiments. An interesting result is that the maximum convection preferentially occurs in the vicinity of the warm pool edge with a local minimum nearby the highest SST. This is consistent with the observed dislocation of the maximum convection and SST in some tropical regions.

2. Numerical Model and Experiment Design

We use the two-dimensional version of the nonhydrostatic Eulerian/semi-Lagrangian anelastic model [Smolarkiewicz and Margolin, 1997]. The zonally oriented 16000 km × 24 km computational domain is on the equator with a 4-km horizontal and 0.3-km vertical grid spacing. Rigid and free-slip vertical boundary conditions are employed at the top and bottom, and at the rigid lateral boundaries the potential temperature and water vapor mixing ratio are set to their respective domain-averaged values at each time step to avoid possibly spurious large gradients [Liu and Moncrieff, 2004a]. A Rayleigh absorber is applied to the uppermost 6 km and the outermost 500 km to reduce the unrealistic reflection of gravity waves.

Without question, a two-dimensional framework has limitations in simulating the convective response to a given SST distribution in the real world. For example, it excludes three-dimensional mesoscale convective systems and especially inherently three-dimensional shear-parallel cloud clusters. The two-dimensional (i.e., zonally oriented) assumption also excludes realistic large-scale tropical waves, dynamical impacts of planetary rotation, and meridional circulations that are potentially important in regulating spatial patterns of convection. However, an advantage of the two-dimensional modeling approach is idealization that facilitates the interpretation of the simulation results without the complication of the aforementioned physical processes. More importantly, the two-dimensional simulations lay a basis for fully three-dimensional simulations that will soon be computationally practicable. In addition, the ability of two-dimensional cloud-system-resolving models in generating large-scale organization of tropical convection has been fully demonstrated in some previous studies [e.g., Oouchi, 1999; Grabowski and Moncrieff, 2001; Liu and Moncrieff, 2004b; Tulich and Mapes, 2008].

Physics processes include warm-rain microphysics [Grabowski and Smolarkiewicz, 1996] and ice microphysics [Grabowski, 1999], subgrid-scale turbulence parameterization [Smagorinsky, 1963], bulk aerodynamic surface flux algorithm [Fairall et al., 1996], and a cloud-interactive radiation transfer scheme [Kiehl et al., 1994]. The diurnal cycle of solar radiation is excluded by assuming a constant zenith angle. The underlying surface is an ocean characteristic of either a relatively broad or narrow warm pool having a background SST of 299 K and a 4 K anomaly (Figure 1), corresponding to a hyper-Gaussian form \( \text{SST}(x) = 299 + 4 \exp(-x^2/L^2) \) and a standard Gaussian form \( \text{SST}(x) = 299 + 4 \exp(-x^2/L^2) \), respectively, where \( L = 2000 \) km is the half-width of the warm pool. These two types of SST distributions were previously applied in the aquaplanet monsoon simulations [Yano and McBride, 1998] and in the mock-Walker circulation simulations [Raymond, 1994], respectively.

The following presentation will concentrate on five 60-day numerical experiments as summarized in Table 1: (1) the broad warm pool (CTL); (2) the narrow warm pool (NWP); (3) the broad warm pool with uniform radiative forcing (URF); (4) the broad warm pool with constant surface wind speed used only for surface flux computation (CSW); and (5) the broad warm pool without surface friction (NSF). In general, the wide-warm-pool SST pattern is relatively more comparable to the reality in the tropics, and therefore is utilized for our control experiment. The narrow-warm-pool experiment is designed to test how the simulated convective pattern relies on the SST distribution and to a lesser degree, to compare with the previous simulation results using the similar SST distribution by Raymond [1994]. The remaining three experiments are intended to explore how some important physical processes, including the wind-induced surface flux variability, horizontally differential radiation and surface friction, affect the simulated convective behavior and attendant large-scale circulations. They are also used to identify the critical factors responsible for the distinct results between the control simulation and the GYM simulation. Additional sensitivity experiments about the impacts of lateral boundary condition, domain size, and SST distribution are performed and will be briefly mentioned.

All simulations are initialized using a single thermodynamic sounding based on the averaged condition over the Intensive Flux Array (IFA) during the 19–26 December 1992 period of TOGA COARE and start from a resting atmospheric state. In addition, small random temperature
and moisture perturbations are added in the lowest one kilometer every 30 minutes during the first day to generate spatial variations in the model fields and trigger convective development.

3. Control Simulation: Wide Warm Pool

3.1. Organized Structures

[12] The control simulation uses the wide warm pool SST distribution and includes full physics. Figure 2a shows the space-time distribution of surface precipitation rate during the 60-day integration. Only the central part of the computational domain is displayed because the remaining portions are convection-free except for the intermittent weak convection near the lateral boundaries. As expected, precipitating convection develops first over the high SSTs after 1–2 days of simulation, presumably due to the relatively stronger surface entropy fluxes associated with the large air-sea thermal contrast therein and the greater convective destabilization than over the low SSTs. Scattered weak convection occurs over the colder SSTs during the early few days, but deep precipitating convection is almost completely suppressed thereafter by the stabilization arising from the compensating descent. Throughout the integration convective activity is mostly confined within a 2000-km wide zone, comparable to the extent of the high SST with roughly a 4 K heating anomaly, and weakens rapidly away from the warm water. The convective distribution is not exactly symmetric with respect to the warm pool. Mesoscale convective organization is often initiated over the warmest water, subsequently propagates slowly toward the colder water at a speed of approximately 2–4 m s$^{-1}$ and dissipates near the edge of the warm pool. This coherent precipitation pattern appears alternatively on each side of the warm pool with a roughly 2-day periodicity. This distinctive transient feature is very likely related to the cyclic buildup of convective available potential energy (CAPE) by destabilization due to radiative cooling and surface processes, and stabilization of active convection [Liu and Moncrieff, 2004a]. Occasionally, convection starts near the warm-pool periphery and moves toward the center. Inspection of the kinematic and thermodynamic fields (not shown) indicates that the sustenance and sluggish movement of organized convective systems are associated with weak surface-based evaporatively generated cold pools and their interaction with convectively induced weak shear.

[13] The temporally averaged precipitation as indicated in Figure 3 covers approximately 3000 km, roughly the warm-water anomaly of at least 3 K, and is not symmetrically distributed with respect to the warmest center. Precipitation peaks in the vicinity of the warm-pool periphery and dramatically decreases toward the colder SST. A local minimum is notable at the center of the warm pool. This spatial pattern is in stark contrast to the robust single maximum at the highest SST reported in previous modeling studies by Raymond [1994] and GYM (see later discussion).

[14] Figure 4 displays the perturbation fields temporally averaged over the last 30 days based on the hourly archived data set. The physical fields are smoothed with a 200-km running-mean filter to eliminate the smaller-scale noise and highlight the organized structures. The horizontal wind in Figure 4a displays a striking symmetry with respect to the warm pool. The wavelike vertical structure of wavelength half of the tropospheric depth is quite similar to the simulation results in GYM and that by Liu and Moncrieff [2004a] and was also observed in natural tropical flow [Johnson et al., 1999]. The airstream in the lowest 2 kilometers flows toward the warm pool and above 9 km goes outward away from the pool, leading to a low-level convergence and an upper-level divergence and thereby significant ascent over the high SST. This convergence-divergence pattern is consistent with enhanced convective activity over the warm water that draws in warm and moist boundary-layer air from both sides of the warm water, transports it upward, and detains it in the upper troposphere. The two branches of strong ascending motion are consistent with the double precipitation peaks (Figure 3) and support the dual weakly coupled thermal circulations, substantially different from the single branch of upward motions centered at the highest SST in previous simulations (GYM). Weak widespread descent dominates over the cold water and is responsible for the suppressed convection therein. The divergent flow around the 4-km level is associated with the shallower convection [Liu and Moncrieff, 2004a], whereas the overlying weak convergence is attributable to the adjacent descending warm and dry air that is unable to reach the lower troposphere and diverges in the middle. This double-cell structure was alternatively interpreted as a consequence of the deviation of the model temperature profile from the climatology by GYM, and the vertical variation of static stability by Mapes [2001].

[15] The temperature perturbation (Figure 4b) shows moderate cooling relative to the initial atmosphere in the lowest two kilometers and significant warming elsewhere. The horizontal distribution is characteristic of relatively a warmer lower and upper troposphere and a colder mid-troposphere over the warm pool compared to the surroundings. The reflection of SST variation is confined to the lowest two kilometers, resulting in baroclinically generated flows toward the warm water at low levels. The extensive mid-level warming over the cold water is obviously a consequence of the adiabatic compensating subsidence warming. The water vapor perturbation (Figure 4c) features moistening throughout the troposphere in the active convection area, and significant drying over the neighboring low SST regions. They result from convective transport and detrainment and compensating subsidence, respectively. The spatial contrast in moisture is much greater than in the temperature.

[16] The pressure distribution (Figure 4d) hydrostatically corresponds to the temperature perturbation and exhibits a relatively low pressure in the lower and upper troposphere.
and a relatively high pressure in between and near the tropopause over the warm pool. A reversed negative-positive perturbation pattern in the vertical is visible on the flanks. The horizontal gradient is generally consistent with the large-scale flow pattern (Figure 4a); namely, an outward (inward) pressure gradient is in correspondence with the zonal wind blowing from warm (cold) water to cold (warm) water. In particular, the strongest pressure gradient near the surface supports the prominent converging airflow toward the warm pool.

3.2. Water Vapor Budget

[17] The water vapor budget quantifies the moist physical processes and precipitation features. In a two-dimensional
anelastic framework, vertically integrated water vapor mixing ratio equation is

\[
\frac{\partial [\rho q_v]}{\partial t} = -\frac{\partial [\rho u q_v]}{\partial x} + E - P + D_{qv} \tag{1}
\]

where \([\ ]\) represents the vertical integration from surface to model top, and the four terms on the right-hand-side represent water vapor convergence, surface evaporation, precipitation, and subgrid diffusion which is assumed to be negligible.

Figure 5 shows that the convectively generated horizontal convergence increases the moisture content (a moisture source) over the convective region and decreases it (a moisture sink) over the broad cloud-free area. The vertical distribution of the water-vapor convergence is primarily confined to the lowest 2.5 kilometers (not shown), associated with the lower-tropospheric convergent flow (Figure 4a). The water vapor convergence is well correlated with the precipitation distribution: the precipitation intensity is proportional to the local water-vapor convergence, and water-vapor divergence consistently appears in non-precipitating areas. On the whole, the surface moisture flux consistently increases from the low-toward-high SST and attains its maximum in the neighborhood of the warm-pool periphery, whereas a local minimum appears at the warm-pool center. As expected, surface evaporation is a water vapor source everywhere, but its peak is located a couple of hundreds kilometers away from the precipitation maximum. Over the cold water, the water-vapor generation from evaporation is approximately balanced by the horizontal transport, whereas over the warm water the water-vapor reduction associated with precipitation is mostly balanced by water vapor convergence. The local evaporation is the secondary moisture source for in situ convection and precipitation. In summary, from the perspective of the moisture budget, the convectively generated horizontal

Figure 3. Spatial distribution of precipitation rate averaged over the 60-day integration in simulation with wide warm pool (thick solid line), narrow warm pool (thin solid line), horizontally uniform radiative heating (dashed line), and constant surface wind speed in calculations of surface fluxes (dotted line). A 200-km running mean filter is applied.

Figure 4. Physical fields averaged from days 30 to 60 for simulation with wide warm pool and full physics. (a) Horizontal wind (2 m s\(^{-1}\) contour interval) and vertical velocity (light and dark shadings represent upward motion greater than 0.5 and 2.5 cm s\(^{-1}\), respectively). (b) Temperature perturbation (1 K contour interval). (c) Water vapor mixing ratio perturbation (1 g kg\(^{-1}\) contour interval). (d) Pressure deviation relative to the horizontal mean (2.5 mb contour interval). (e) Radiative heating (0.5 K day\(^{-1}\) contour interval) and condensate (light-to-dark shadings progressively represent condensate greater than 0.001, 0.01, and 0.5 g kg\(^{-1}\), respectively). A 200-km running mean filter is applied.
moisture convergence dominates local evaporation as the primary source that balances moisture loss by precipitation.

4. Sensitivity Experiments

Additional experiments are performed to identify the critical factors controlling the simulated convection and attendant large-scale organized flow patterns.

4.1. Narrow Warm Pool

The narrow warm-pool (Figure 1) experiment examines the impact of warm-pool morphology on the spatial distribution of convective activity and precipitation. The Hovmoller diagram (Figure 2b) indicates that the space-time precipitation pattern is largely comparable to the wide warm-pool counterpart (Figure 2a). Prevalent coherent convective events travel away from the highest SST and are intermittent with weak oppositely traveling organization. The temporally averaged precipitation (Figure 3) also exhibits a similar dual-maximum, although the precipitation zone is slightly narrower than the control simulation, consistent with the narrower warm pool. The domain-mean rainfall amount is almost identical. As expected, the time-mean circulation is comparable to the warm-pool counterpart because of the very similar convective pattern in the two simulations. These results indicate that the double-maximum precipitation around the warm-pool edges and the attendant double-branched circulations are robust and insensitive to the warm-pool extent. Note that a single precipitation peak and ascending shaft centered at the highest SST were produced in previous coarse-resolution simulations using a similar narrow warm-pool SST pattern [Raymond, 1994] (more discussion later).

4.2. Uniform Radiative Forcing

The simulated convection is regulated by cloud-radiative interaction and surface energy fluxes. The moisture uptake from underlying ocean provides the water-vapor source for convective development. On the other hand, cloud-radiative interaction affects the atmospheric environment through multiple processes [e.g., Tao et al., 1996; Liu and Moncrieff, 1998]: such as (1) large-scale destabilization associated with radiative cooling over the whole domain; (2) enhancement of convergence into cloud systems associated with the horizontally differential cooling between cloudy and clear regions; and (3) thermal stratification alteration due to the differential cooling between cloud base and cloud top. Among these radiative impacts, differential heating/cooling induced by spatial variability of water substance is of particular interest because many studies showed that it is an important driver of various types of cloud systems and large-scale circulations [e.g., Gray and Jacobsen, 1977; Cox and Griffith, 1979a, 1979b; Cohen and Frank, 1987; Dudhia, 1989; Slingo and Slingo, 1988, 1991; Raymond, 2000] and, moreover, there exist substantial variations of water vapor and condensate amount between the upward and downward branches in the control experiment (Figures 4c and 4e).

Figure 4e presents the temporally averaged radiative heating/cooling rate in the control simulation. A widespread cooling associated with long-wave radiation prevails over the entire domain, acting to destabilize the large-scale environment. There is significant horizontal variability at each level. A weak cooling less than $-0.5 \text{ K day}^{-1}$ or even a weak heating, presumably related to cloud-base warming [Tao et al., 1996], occurs in most of the lower atmosphere over the warm pool, whereas a comparatively stronger cooling ($-1.5$–$2.5 \text{ K day}^{-1}$) appears in the surrounding cloud-free atmosphere. As a consequence, the radiative temperature tendency over the warm water is generally greater than over the neighboring cold water roughly below 10 km. An opposite, albeit moderate, tendency occurs aloft presumably due to the large long-wave radiative cooling at the cloud top. This spatially differential radiative cooling/heating pattern is broadly consistent with the spatial distribution of atmospheric moisture and cloudiness, and it is anticipated to enhance the low-level convergence and upper-level divergence over the high SSTs and thus influence the intensity and spatial distribution of convective organization.

In order to quantify the potential impact of the inhomogeneous radiative cooling on simulated convective organization, we performed a sensitivity experiment with horizontally uniform radiative tendencies that are the domain averages of the potential temperature tendencies from the radiative transfer model at each time step. The time-space distribution of surface precipitation rate in Figure 2c

![Figure 5. Spatial distribution of surface precipitation (solid line), surface evaporation (dotted line), and water vapor convergence (dashed line) averaged from days 30 to 60 for simulation with wide warm pool and full physics.](image-url)
evinces marked changes when the spatial variability in radiative cooling is eliminated. The foremost is the much wider convective zone. Unlike the control simulation, coherent convective patterns commonly initiate near the edge of the warm water, and less organized convection is prevalent over the wide-warm pool. The propagating systems are weaker and less persistent and travel more slowly compared to the control simulation. Nevertheless, the temporally averaged precipitation still features double peaks, although the maxima are displaced hundreds of kilometers farther off the highest SST (Figure 3). The domain-mean rainfall rate is also reduced by roughly 10% when the cloud-interactive radiative cooling is removed. These results illustrate that the cloudy—clear differential radiative heating intensifies convection and confines convective activity toward the warmest ocean by enhancing the SST-gradient-driven circulation. This finding is physically understandable: as aforementioned the net radiative cooling in the upward branch is weaker (greater) at low levels (upper levels) than in the neighboring downward branches, leading to a positive feedback on existing convection over the warm pool by intensification of low-level convergence and upper-level divergence. As a consequence, convective activity would be stronger in the presence of cloud-radiative interaction than without it. Moreover, enhanced convection implies that convective zone would be narrower or more concentrated if the available water vapor source does not change much.

The exclusion of cloud-interactive radiation also modifies the time-mean flows as shown in Figure 6. The velocity field (Figure 6a) demonstrates two completely decoupled circulations, each having narrow ascent in the neighborhood of the warm-pool periphery and wide compensating sinking motion over the low SST. However, the dual-cell organization in the vertical is evident as in the control simulation. On the basis of inflow-outflow strength, the lower-tropospheric cell is weakened but the upper tropospheric one is intensified. The temperature and water-vapor mixing ratio perturbations (Figures 6b and 6c) are considerably weak over the warm-pool water, compared to the cloud-interactive simulation (Figures 4b and 4c). The adiabatically induced warm and dry deviations over cold water are decreased, in agreement with the weakened large-scale circulations. In the pressure field (Figure 6d), the reduced low-level and enhanced upper-level gradients as well as the broadened weak horizontal gradients over high SSTs are consistent with the circulation differences between the two simulations.

4.3. Constant Surface Wind Speed

There are considerable spatial variabilities in air-sea heat and moisture exchanges (Figure 5). As well as SST, surface fluxes depend on surface wind speed, surface air temperature and moisture. The inspection of their spatial distributions indicates that the surface-flux distribution is primarily modulated by the spatial variability in surface wind speed; especially, the local extreme fluxes well correspond with the extreme surface wind over the warm waters (Figure 7a). In contrast, the surface air temperature and moisture roughly track the imposed spatial variations in SST (Figures 7b and 7c). As a consequence, the relatively warmer and moister surface air over the warm pool tend to offset the positive impact of higher SSTs on energy uptake therein. Because the surface moisture transport is the only water-vapor source for convection, we hypothesize that surface flux variation associated with the differential surface wind speed plays an important role in determining convection, and precipitation strength and distribution.

To test the hypothesis, a sensitivity experiment is conducted with a time-invariant and horizontally uniform surface wind speed of 6 m s\(^{-1}\) applied only in the computation of surface fluxes. The specified wind speed is approximately the space and time-averaged value in the control simulation. This setup eliminates the wind-induced
sea-air heat exchange (WISHE) mechanism [Emanuel, 1987; Neelin et al., 1987] which was found essential to the development and maintenance of the equatorial maximum convection in two-dimensional numerical modeling of convection on an equatorial β plane [Liu and Moncrieff, 2004a]. Although many of convective characteristics are retained, there exist significant differences compared to the control experiment. In particular, the convective activity and resultant precipitation are confined to a much smaller region (roughly 1000 km wide) after about one week into the integration (Figure 2d). The region of concentrated convection undergoes a periodic transition around the highest SST. As in other simulations, the dual-maximum of convection occurs, but the precipitation is less symmetric with regard to the SST distribution (Figure 3). The peak on the left side is stronger than on the right side, and the minimum is off the warm pool center.

In summary, the most noticeable change is the contraction of precipitation area when the wind-induced surface flux variations are excluded. In other words, the non-uniform energy fluxes resulted from the varying surface wind broaden the convective region toward the warm-pool periphery. This result is physically explained as follows. The surface wind speed in CTL shows a local minimum at the warm-pool center and a double maximum at the edges as indicated in Figure 7a. As a result, the surface flux would be artificially increased (decreased) around the central warm pool (peripheries) when a uniform surface wind is applied in the surface-flux calculation. Obviously, this kind of surface-flux modification is beneficial to convection at the warm-pool center but harmful to convection at the warm-pool edges, leading to more concentrated convective activity over the warmest SST.

4.4. Surface Friction

[28] Organized tropical convection and the attendant large-scale flow that develop in response to prescribed SST gradients have been numerically investigated in several previous studies. Raymond [1994] performed such simulations using a hydrostatic, two-dimensional, non-rotating model with very simple formulations of cloud, precipitation and radiation processes. The results showed the Walker-like circulation with a unicellular ascending branch positioning at the central warm pool. The corresponding rainfall inten-
sity closely follows the SST distribution and peaks at the warmest ocean, in sharp contrast with the spatial pattern in our simulations. This difference is perhaps not surprising in view of the very coarse grid resolution (500 km) and highly simplified treatment of convective and radiative processes in Raymond’s experimental setting.

[29] GYM reported two-dimensional cloud-system-resolving simulations of this type of problem with the same numerical model as in our study. Three simulations were discussed in detail, corresponding to fully interactive radiation, noninteractive radiation with a prescribed constant radiative cooling rate, and adjusted radiation. (Their third simulation applies interactive radiation in conjunction with an adjustment procedure to ensure that the horizontally averaged temperature tendency matches the tendency in the noninteractive radiation simulation.) As expected, there are many features in common in the time-mean flow and rainfall between GYM and our simulations, such as the upward motion and active convection over the warm ocean and the cloud-free downward motion over the cold ocean, and the double-cell vertical structure. Nevertheless, there are some differences, for example, the single convection maximum and associated single ascending branch centered at the highest SST. This property also appeared in

Figure 8. Precipitation rate Hovemoller (x–t) diagram for simulation without surface friction. The blue and red shadings represent rain rate greater than 1 and 10 mm h⁻¹, respectively.

Figure 9. Spatial distribution of precipitation rate averaged over the 60-day integration in simulations with (solid line) and without (dashed line) surface friction.

Figure 10. Physical fields averaged from days 30 to 60 for simulation without surface friction. (a) Horizontal wind (2 m s⁻¹ contour interval) and vertical velocity (light and dark shadings represent upward motion greater than 0.5 and 2.5 cm s⁻¹, respectively). (b) Temperature perturbation (1 K contour interval). (c) Water vapor mixing ratio perturbation (1 g kg⁻¹ contour interval). (d) Pressure deviation relative to the horizontal mean (2.5 mb contour interval).
Raymond’s simulations with parameterized convection, as opposed to the dual precipitation peaks and decoupled ascending branches off the warm-pool center. In order to examine what aspect of the experiment setup is responsible for these differences, further sensitivity experiments examined the impacts of lateral boundary conditions, SST distribution, computational domain size, and surface friction. We found that the lack of surface friction in GYM mostly accounts for the aforementioned differences from our simulations. The differences in boundary conditions, SST and domain size were of secondary importance.

Figure 8 indicates that the convective activity is concentrated over a narrow region of the warmest ocean during most of the friction-free simulation. In particular, the almost uninterrupted precipitation roughly between days 12–35 and after day 53 indicates a persistent, unicellular, saturated updraft in the center of the domain. As a result, the temporally averaged rainfall in Figure 9 has a sharp peak at the warmest SST. Note that in comparison with the control simulation, the accumulated precipitation, albeit quite narrowly distributed, is stronger due to the enhanced surface fluxes associated with the greater wind speed in the absence of surface friction.

Figure 10 shows the time-averaged fields of horizontal and vertical velocities, and temperature, water vapor mixing ratio and pressure perturbations to further evaluate the influence of surface friction. Many aspects of the horizontal flow are comparable to the control simulation and GYM. The large-scale circulation is characteristic of a vertical structure having two inflow regions, the stronger one at the surface and the weaker one at 8 km, and two outflow regions, the stronger one in the upper troposphere above 9 km and the weaker one at 4 km. The major difference from the control simulation is the concentrated convergence/divergence over the warm pool, leading to a strong unicellular updraft as in GYM and that of Raymond [1994]. Another salient difference is the more intensive and extensive horizontal wind speed, implying a stronger Walker-type circulation in the friction-free experiment. The temperature and water-vapor perturbations are very similar to their respective counterparts in the control simulation except that the warm and wet anomalies have local peaks at the highest SST. The surface friction reduces pressure perturbations in the lower troposphere, but increases the horizontal gradients near the center of the domain.

The preceding comparisons suggest that the surface friction has important effects on the simulated convective organization and large-scale circulation. As expected, its inclusion weakens the overall circulation and convection. More interestingly, friction influences the horizontal scale of convection. It displaces the convective activity away from the warmest water and thus leading to two precipitation maxima and two weakly coupled ascending branches near the warm-pool periphery. It is the primary physical process that accounts for the differences between our simulation results and the GYM results. However, the physical interpretation of the significant frictional effect needs further investigation.

5. Conclusions

The intensity and spatial pattern of tropical convection are strongly controlled by SST distributions. We
examined organized convective features and the large-scale flow response to idealized SST distribution by using two-dimensional cloud-system-resolving simulations. We quantified the physical processes that influence the strength and horizontal extent of convection. Figure 11 schematically illustrates the spatial patterns of deep precipitating convection and the corresponding mean airflow structures. The full-physics experiment (Figure 11a) is characteristic of double loosely coupled circulation cells in the horizontal, whose upward branches are well coupled with active convection near the warm-pool peripheries. The exclusion of heterogeneous radiative forcing (Figure 11b) displaces convective activity away from the warm water and leads to dual completely decoupled circulations, whereas the exclusion of the surface friction (Figure 11c) confines moist convection toward the highest SST and results in a large-scale single-cell morphology. All cases exhibit a complex vertical structure with two somewhat decoupled circulations in the lower and upper troposphere. The major findings are summarized as follows.

[34] 1. Convection intensity and precipitation amount tend to peak around the edge of warm pools, and local minimum occurs near the highest SST. This result may explain why the observed maximum convection often does not collocate with the warmest SST.

[35] 2. The strong horizontal contrast in radiative heating/cooling associated with cloud-radiative interactions between cloudy and convection-free areas strengthens convection and, more importantly, concentrates convective activity toward the highest SST.

[36] 3. Conversely, the non-uniform energy fluxes associated with the spatial variability of surface wind speed broaden the active convection toward the colder water.

[37] 4. The surface friction drastically impacts convective spatial variability, displacing the maximum convection away from the warm-pool center and causing dual circulation cells.

[38] 5. These results suggest that as well as the horizontal temperature/pressure gradients transmitted from the underlying ocean [Lindzen and Nigam, 1987], other physical processes cannot be neglected in the interpretations of observed tropical convective features: Cloud-interactive radiation, surface friction, and spatial variability of surface gustiness are important to the organization of convection over the warm pools.

[39] The above conclusions are subjective to the use of a two-dimensional modeling framework. Nevertheless, this study is helpful in understanding and explaining observed behavior of SST-convection correlation in the tropics. It also provides a basis for future three-dimensional modeling studies and investigations of convective organization in more complicated and observed SST distributions.

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