Cloud Resolving Modeling of Tropical Circulations Driven by Large-Scale SST Gradients

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ABSTRACT
This paper considers interactions between the moist atmospheric convection and the large-scale flow driven by the large-scale gradient of sea surface temperature. A two-dimensional computational framework is used with the horizontal domain size of 4000 km in which both the convective dynamics and the large-scale flow are resolved. Rotational effects are not considered. Simulations are performed using either a prescribed temperature tendency mimicking the effects of radiative processes or a fully interactive radiation transfer model. The simulations are performed for a period of 60 days with quasi-equilibrium conditions attained after about a month.

The time-mean large-scale flow in the simulations features an ascending branch occupied by moist convection over a warm ocean and a cloud-free descending branch over a cold ocean. The time-mean flow for the prescribed radiation case features a complex vertical structure characterized by two somewhat decoupled circulations in the lower and upper troposphere. This is in stark contrast with the predominant first-baroclinic-mode structure typical of the observed large-scale tropical circulations, which is characterized by a single cell. An idealized dry model featuring prescribed convective heat source suggests that the complex vertical structure is directly related to the deviation of the model temperature profile from the climatology. Quasi-two-day oscillations are a major transient feature of the simulations. The oscillations are associated with radiation of gravity waves from the convective branch into the descending branch.

Inclusion of the interactive radiation results in a significant modification of the large-scale flow and has a dramatic impact on the strength and horizontal extent of convection. Water vapor and cloud condensate strongly interact with radiative processes to induce these paramount effects on the tropical large-scale circulations.

1. Introduction
The Tropics cover a significant part of the earth’s surface and play an essential role in the climate system. Yet the dynamics of the tropical atmosphere is poorly understood when compared to the dynamics of the mid-latitudes. This is because in the Tropics—unlike in the midlatitudes where rotational effects dominate—the large-scale dynamics depends critically on the diabatic processes through which the atmosphere exchanges energy with the underlying surface and with space. On timescales relevant to climate, dynamics of the tropical atmosphere reflects an approximate balance between energy loss due to radiative processes and energy gain due to heat fluxes from the ocean surface. Most of the energy gained from the surface is in the form of the latent heat of water vapor. The latent energy is released within tropical convective clouds with water returning to the ocean surface as rain. In turn, the moist convection affects the exchange of heat, water, and momentum between the atmosphere and the ocean and has an essential impact on solar and terrestrial radiative fluxes. Consequently, tropical dynamics and the attendant hydrologic cycle span a wide range of spatial scales, from hundreds of meters (e.g., boundary layer eddies, shallow convection) to thousands of kilometers (e.g., equatorially trapped waves, monsoon circulations). With present...
computer power, only a fraction of this range can be simultaneously resolved in a numerical model.

Traditionally, two distinct computational approaches have been used to study various aspects of the tropical dynamics. The first considers large-scale aspects by applying a large-scale numerical model with a horizontal resolution of the order of 100 km. In such an approach, the effects of moist convection can be included only through parameterization because the horizontal resolution of such models is much too coarse to represent the convective dynamics explicitly. Numerous idealized studies have been performed to study the large-scale tropical dynamics with this type of setup assuming that the earth is totally covered by the ocean, as pioneered by Hayashi and Sumi (1986). However, the results of such experiments are usually sensitive to the choice of cumulus parameterization (see Yano et al. 1998 and the references therein).

The alternative approach considers the details of convective dynamics on timescales relevant for the tropical climate by applying a numerical model with a relatively high spatial resolution (∼1 km). In such a model, the phase changes of water interact directly with convective dynamics, and the effects of clouds on radiative fluxes and surface processes are thereby more realistic. Unfortunately, the horizontal extent of the domain in high-resolution (or cloud resolving) models is usually too small (a few hundred kilometers on a side at most) to allow for large-scale dynamical feedback (e.g., Grabowski et al. 1996b, 1998; Wu et al. 1998, among others).

There is no doubt, however, that the ultimate approach to the tropical dynamics has to involve both the large-scale dynamics and the cloud-scale dynamics represented within a single dynamical framework. Because of the necessary high horizontal resolution, this consistent approach is currently possible only in a context of a two-dimensional (2D) numerical model. This paper presents 2D cloud resolving simulations in which a large-scale atmospheric circulation is driven by a prescribed sea surface temperature (SST) distribution.

A similar problem was recently considered by Raymond (1994), who considered a 2D large-scale flow driven by the SST distribution using parameterized convection. The resulting large-scale circulation was referred to as the “mock Walker circulation.” We move one step further in the present study. Instead of parameterizing convective processes in a hydrostatic large-scale model, we explicitly predict them in a nonhydrostatic cloud resolving model. For the most part, similar large-scale flows develop in the present simulations as in Raymond (1994). However, the emphasis of the present paper is on the interaction between the large-scale flow and convection, and no attempt will be made to compare our results with those of Raymond (1994) in detail. It should also be mentioned that the cloud resolving simulations reported herein follow a pilot simulation presented in section 5 of Grabowski (1998, hereafter G98).

The paper is organized as follows. Section 2 presents the numerical model and provides details of the simulations performed. The general features of the model results are discussed in section 3. Section 4 focuses on the quasi-equilibrium mean circulation. The transient model behavior is discussed in section 5 with the focus on quasi-two-day oscillations. Model results and their relevance to the natural large-scale circulations are discussed in section 6.

2. Numerical simulations

The cloud resolving simulations described in this paper consider a 2D convecting large-scale flow that develops in a periodic horizontal domain with prescribed horizontal SST distribution. The numerical model applied herein is a massively parallel version (Anderson et al. 1997) of the two-time-level, nonhydrostatic Eulerian/semi-Lagrangian anelastic fluid model of Smolarkiewicz and Margolin (1997) with the moist precipitating thermodynamics applied as discussed in Grabowski and Smolarkiewicz (1996) and in G98. The Eulerian version of the model is used.

In the simulations, the 4000-km-long and 25-km-deep domain is covered with a regular grid with about 1.8-km resolution in the horizontal direction and 1/3 km resolution in the vertical. The time step is 15 s. A gravity wave absorber is applied in the uppermost 8 km of the domain. The domain is periodic in the horizontal direction, and the SST varies according to a sine function with 28°C in the center of the domain and 24°C at the periodic lateral boundaries. Neither rotational effects nor surface friction are considered.

Three numerical simulations have been selected for detailed analysis. In addition, a few additional simulations will be mentioned to support conclusions drawn from the analysis of the three primary simulations. An important modification of the pilot simulation presented in G98 is an application of the Tropical Ocean Global Atmosphere Coupled Ocean–Atmosphere Response Experiment (TOGA COARE) surface flux algorithm of Fairall et al. (1996). However, for simplicity, the coolskin and warm-layer corrections are omitted. Because the simulated large-scale flow is characterized by weak surface winds (between 1 and 3 m s−1), the simple surface flux algorithm applied in G98 led to unrealistically large differences between the surface air temperature and the SST. In general, in this type of simulation, the energy source associated with the sensible and latent surface heat fluxes has to balance the energy loss due to radiative processes after equilibrium conditions are achieved. It follows that the time- and space-averaged surface fluxes are directly tied to the radiative temperature tendency. Given the radiative temperature tendency and the surface winds, the model produces the required value of the total surface flux by adjusting the
air–sea temperature difference. After incorporating the TOGA COARE surface flux algorithm, the air–sea temperature difference in the simulations presented in this paper is typically between 2 and 5 K, which is a more reasonable value compared to the values in excess of 10 K present in some parts of the domain in the pilot simulation of G98.

The three simulations discussed in this paper differ only in the way radiative processes are represented. The first simulation, referred to as prescribed radiation (PR), prescribes a horizontally homogeneous temperature tendency profile mimicking the mean (i.e., noninteractive) effect of radiative cooling. The temperature tendency profile assumed corresponds to a 1.5 K day$^{-1}$ sink of the potential temperature at all tropospheric levels as in G98. The second simulation represents interactive radiation (IR) by applying the National Center for Atmospheric Research (NCAR) Community Climate Model radiation code (Kiehl et al. 1994). However, to be consistent with the simulation PR, a diurnal cycle is not considered for the solar radiation. Instead, the solar constant is reduced to 436 W m$^{-2}$ (i.e., the nominal solar constant divided by $\pi$) and a zero zenith angle is assumed. The radiative calculations are performed every 15 min of model time.

Furthermore, an additional simulation is performed with a setup intermediate between the two previous cases because application of the radiative transfer model (IR) results in averaged radiative tendencies substantially different from those applied in the PR simulation. This third simulation, referred to as adjusted radiation (AR) applies interactive radiation combined with an adjustment procedure, which ensures that the horizontally averaged temperature tendency at any given level matches the tendency prescribed in the PR simulation. The adjustment procedure is given by

$$Q_A(x, z, t) = Q_P(x, z, t) - \langle Q_P(x, z, t) \rangle + Q_R(z),$$  

where $Q_R$ is the potential temperature tendency due to radiative processes derived from the radiation transfer model, $\langle \cdot \rangle$ represents the horizontal average over the entire computational domain at a given height, and the superscripts $A$ and $P$ stand for the adjusted and prescribed radiative temperature tendencies, that is, those applied in AR and PR simulations, respectively. It follows from (1) that the AR simulation applies the same domain-averaged radiative temperature tendencies as PR, but the differences between clear and cloudy columns, as well as columns with high and low water vapor content, are preserved. The purpose of the AR simulation is to shed light on the origin of the dramatic differences between PR and IR simulations.

The model output is archived every 3 h for all three simulations and the synthetic datasets are used in the analyses.

### 3. Approach to the quasi-equilibrium regimes

The simulations are initiated using a 0000 UTC 1 September 1974 Global Atmospheric Research Program (GARP) Atlantic Tropical Experiment (GATE) sounding (as in Grabowski et al. 1996b) to each model column of an atmosphere at rest. Because of the destabilization caused by radiative and surface processes, convection develops during the first day of the simulations. The large-scale circulation is established within the next few days. It is characterized by the ascending branch featuring deep convection over the warm ocean and the descending clear-sky branch over the cold ocean. The simulations are run until the equilibrium conditions (in terms of the domain-integrated energy and water) are established. Because of the slow descent outside the convecting part of the domain, the adjustment toward equilibrium takes a few weeks (cf. G98, see also Tompkins and Craig 1998b).

Figure 1 shows the domain-averaged temperature budget for the 60-day-long simulations of the PR and IR cases, that is, simulations applying prescribed and interactive radiation. Moisture budgets are not shown; however, the moisture budget for the PR case is similar to the one shown in G98 (Fig. 14 therein). The AR simulation (not shown) displays features found in both PR and IR simulations. The figure shows that the system attains the equilibrium state in about 3 weeks. The quasi-equilibrium regimes could be deduced from the last 40 days of the simulations; however, only the last 15 days will be used herein. The equilibrium density-weighted temperature and the precipitable water are quite different for all three simulations, namely, 261 K and 28 kg m$^{-2}$ for PR, 265 K and 23 kg m$^{-2}$ for AR, and 267 K and 31 kg m$^{-2}$ for IR.

The most striking feature of the temperature and moisture budgets, noted in G98, are the quasi-two-day oscillations in the temperature and the precipitable water (not shown), especially in the PR simulation, as well as the temperature and moisture sources. These oscillations are most regular in the AR simulation (not shown) and disappear in the last week of the PR simulation. An explanation of these oscillations will be presented in section 5.

Figure 2 shows the spatial distribution of the temperature tendency due to radiative flux divergence $Q_{\theta} T_{\theta}$ (where $T_{\theta}$ and $\theta$ are ambient temperature and potential temperature profiles of the anelastic system) and its domain-averaged vertical profiles for the last 15 days of AR and IR simulations. The domain-averaged profile for the AR simulation is identical to the profile applied across the entire domain in the PR simulation. The prescribed and interactive radiative temperature tendency profiles differ considerably. The interactive profile features more cooling in the lower half of the troposphere and less cooling in the upper half. However, the density-weighted integrals of the radiative tendency, $\int \rho Q_{\theta} T_{\theta} \, dz$ (i.e., the source of the density-weighted
temperature or dry enthalpy), are almost the same with prescribed and interactive radiation (see Figs. 1a and 1b). This further implies that the total convective heating rate for the whole system is also approximately invariant to a change of the radiation scheme because it balances the radiative cooling for the whole system in equilibrium. The spatial distribution of the radiative cooling in the IR simulation is to a large degree defined by the distribution of the water vapor and cloud condensate (see Fig. 4 below). In particular, the layer of strong cooling sloping toward the cold SSTs in midtroposphere coincides with the strong local decrease of moisture with height. The positive radiative tendency in the middle of the domain is presumably related to the distribution of condensed water (i.e., cloud systems).

Figure 3 shows the Hovmöller ($x$–$t$) diagrams of the surface precipitation rate to demonstrate evolution and organization of convection for PR, AR, and IR simulations. Initially, in each case the convection covers about a half of the domain, but its subsequent evolution differs substantially. In the simulation in which the assumed profile of radiative cooling is applied homogeneously across the domain (PR, Fig. 3a), precipitating convection covers about one-third of the domain over the warmest SST. A closer inspection shows that various organizations of convection exist in addition to the quasi-two-day oscillations in the domain-averaged variables. After 20 days of the simulation, the convection appears to be organized into convective systems that originate near the center of the domain and travel toward the colder SSTs. However, this organization disintegrates into many smaller convective systems during the last week of the simulation. The relationship of these transient features to the quasi-two-day oscillations is discussed further in section 5.

The organization of convection is drastically different when interactive radiation is applied (IR simulation, Fig. 3b). In this case, convection occupies a narrow zone in the center of the domain. Convective systems are evidently generated outside of this zone and travel toward the center of the domain. The zone with surface precipitation is organized on a scale of just a few hundred
Fig. 3. Hovmöller (x-t) diagrams of the surface precipitation rate for the (a) PR, (b) IR, and (c) AR simulations. Precipitation intensity larger than 0.2 and 5 mm h⁻¹ is shown using light and dark shading, respectively.
kilometers. The precipitating region in the center of the domain is dominated by a slow quasi-hydrostatic saturated ascent (with an updraft speed less than 1 m s\(^{-1}\)) punctuated by weak convective updrafts of 1–2 m s\(^{-1}\).

The difference between the PR and IR cases can be attributed either to the difference of the mean radiative tendency profile or to the distribution of the radiative tendency in the horizontal at a given level. The latter factor is associated with the feedback of moist convection onto radiation. The importance of these two factors is highlighted in the simulation AR (Fig. 3c) where convection tends to occupy a narrow region in the center of the domain as in the IR simulation. This implies that the collapse of the convective region into a narrow band in the IR simulation is a consequence of the gradient in radiative tendency between ascending and descending branches. This mechanism seems similar to that operating on much smaller scales in simulations described in Tompkins and Craig (1998a). On the other hand, convective updrafts in the AR simulation are almost as strong as in the PR. As expected, the strength of convection seems determined by the horizontally averaged radiative tendency. In addition, the quasi-two-day oscillations are very pronounced and regular in the AR simulation. The tendency of the convective region to meander with about a 40-day timescale in the AR simulation is also worth pointing out.

To illustrate main features of the circulation in the quasi-equilibrium regime, Fig. 4 shows the snapshots of the thermodynamic fields for the PR and IR simulations at day 53.00, near the end of the simulations. The main common features (see also Fig. 12 in G98) include a small temperature difference between ascending and descending branches, and a large moisture gradient between the two branches. A strong decrease of moisture with height within a midtropospheric zone sloping from the center of the domain toward the colder SSTs is also noted. This sharp moisture gradient contributes to the strong radiative cooling rate in Fig. 3. However, there are significant differences between the PR and IR simulations. For instance, the IR simulation (Fig. 4b) features a stronger midtropospheric moisture inversion. It also seems to have a better organized boundary layer with pronounced temperature and moisture inversions, which slope upward from cold to warm SSTs. Except for the width of the convective region, the simulation AR is similar to PR and is not shown.

Time- and space-averaged temperature and relative humidity profiles within ascending and descending branches of the large-scale circulation summarize the simulated thermodynamical profiles. The profiles are calculated as averages over 100-km zones because convection occupies only the narrow zone in the middle of the domain in AR and IR simulations. Recall that the model domain extends from \(-2000\) to \(2000\) km with the highest SST in the center, the temperature profiles \(\theta_a\) and \(\theta_d\) in the ascending and descending branches, respectively, are calculated as [cf. (22) in G98]...
FIG. 5. Profiles of the potential temperature and the relative humidity for ascending (solid line) and descending (long-dashed line) branches of the large-scale circulation as defined in text for the (a) PR and (b) IR simulations. The short-dashed line shows the initial profiles based on the 0000 UTC 1 Sep 1974 GATE sounding.

\[ \theta_a(z) = \frac{1}{100 \text{ km}} \int_{-50 \text{ km}}^{50 \text{ km}} \theta(x, z) \, dx \quad \text{and} \quad (2a) \]

\[ \theta_d(z) = \frac{1}{50 \text{ km}} \int_{-1950 \text{ km}}^{-2000 \text{ km}} \theta(x, z) \, dx \]

\[ \quad + \frac{1}{50 \text{ km}} \int_{1950 \text{ km}}^{2000 \text{ km}} \theta(x, z) \, dx. \quad (2b) \]

Analogous expressions are used for the relative humidity. The profiles in (2) averaged over the last 15 days are shown in Figs. 5a and 5b for the PR and IR simulations, respectively. The AR profiles are intermediate between the PR and IR limits and are not shown. Figure 5 also shows the initial temperature and relative humidity profiles, which are based on the 0000 UTC 1 September 1974 GATE sounding.

The temperature profiles differ little between ascending and descending branches of the large-scale circulation, which is characteristic of the tropical atmosphere. In the PR simulation (Fig. 5a), the temperature profiles and the GATE profile in the lower half of the troposphere match quite well. In the upper half, on the other hand, the temperature profiles are much colder than the GATE profile, by up to 20 K at 15 km. The very cold upper troposphere results in high values of convective available potential energy (CAPE) (around 3000 J kg\(^{-1}\)) for a pseudoadiabatic parcel rising from the ocean surface and correspondingly strong convective updrafts (up to 15 m s\(^{-1}\)) in the ascending branch of the large-scale circulation. The IR relative humidity profile, on the other hand, is close to water saturation in the lowest 6 m, which is consistent with the slow saturated ascent in the center of the domain.

Figures 6, 7, and 8 show the time- and space-averaged fields of the virtual potential temperature perturbations from the horizontal average at a given level, and the corresponding vertical and horizontal velocities for the PR, IR, and AR simulations. The fields were spatially smoothed over a distance of 400 km and smoothed fields...
were averaged over the last 15 days of the simulations. For the simulation with prescribed radiative tendency (PR, Fig. 6), the temperature perturbation field shows positive perturbations (with magnitude up to 1 K) in the upper half of the troposphere over warm SSTs and the lower half of the troposphere over cold SSTs (cf. Raymond 1994, Fig. 10). The mean vertical velocity reaches its extreme values in the upper troposphere (about 5 cm s$^{-1}$ maximum and $-2$ cm s$^{-1}$ minimum). A distinctive feature is the upper-tropospheric outflow from the warm SST region with deep convection and two inflow regions, one at the surface and the other at about 7 km. Because the boundary layer temperature basically reflects the SST gradients (see Fig. 5), the flow near the surface seems analogous to the flow considered by Lindzen and Nigam (1987). However, its vertical scale (about 1 km) is much shallower than that assumed by Lindzen and Nigam (i.e., 3 km). There is also a hint of the lower-tropospheric outflow from the region of convection, although it does not show in the time-averaged pattern. Consequently, one can argue that the large-scale flow is represented by two circulation cells: one in the mid-to upper troposphere, and the other in the lower troposphere. The upper cell has larger vertical velocities. The weak upper-tropospheric stability (cf. Fig. 5) requires a stronger subsidence than in the lower troposphere to compensate the prescribed radiative cooling. Such a double circulation pattern is in stark contrast to the single circulation cell characterized by the upper-

![Fig. 6](image1)

![Fig. 7](image2)

![Fig. 8](image3)
The vertical velocity pattern in the interactive radiation simulation IR (Fig. 7) is quite different from PR (Fig. 6). In the IR case, the flow in the upper half of the troposphere is evidently more decoupled from the lower-tropospheric flow than in PR. This is supported by values of CAPE in the IR simulation being too small to support deep convection. By comparing PR and IR simulations, one can argue that the interaction of radiation with clouds and with water vapor plays a key role in shaping the large-scale circulations in the Tropics. The AR simulation (Fig. 8), on the other hand, displays features found in both the PR and IR simulations (e.g., stronger upper-tropospheric circulation cell as in PR, or temperature perturbation distribution resembling IR).

Finally, for completeness, Figs. 9 and 10 show the spatial distributions of time-averaged surface sensible and latent heat fluxes for PR and IR. (The AR has a pattern similar to IR and is not shown.) The magnitude of the surface fluxes varies considerably with time due to the quasi-two-day oscillations (cf. Fig. 1), but the spatial distribution remains approximately the same (except for the areas directly affected by convection). The spatial distribution of surface fluxes is considerably more complex than in the case of parameterized convection in Raymond (1994, Fig. 8). The total domain-averaged surface fluxes (i.e., sums of sensible and latent fluxes) are approximately equal in the PR and IR simulations, which is consistent with the similar radiative energy sinks as illustrated in Fig. 1.

In sections 4 and 5, two aspects of the model results, namely, the mean large-scale circulation and the quasi-two-day oscillations, are examined in more detail.

4. The quasi-equilibrium large-scale circulation

a. The temperature profiles

The evolution of the domain-mean potential temperature profile is deduced by horizontally averaging the temperature equation over the horizontally periodic computational domain (cf. Sui et al. 1994, section 2b; Grabowski et al. 1996a, section 4):

$$\frac{\partial(\theta)}{\partial t} = -\frac{1}{\rho_o} \frac{\partial (\rho_u w' \theta' \beta)}{\partial z} + \langle Q_s \rangle$$

$$+ \frac{\theta_o L_v}{T_v c_p} (c + d - e - s) + \frac{1}{\rho_o} \frac{\partial (F_q)}{\partial z}, \quad (3)$$

where $\rho_o(z)$ is the base-state anelastic density profile; primed variables represent deviations from the horizontal mean; $L_v$ and $c_p$ are the latent heat of vaporization and specific heat at constant pressure; $c$, $d$, $e$, and $s$ stand for rates of condensation, deposition, evaporation, and sublimation, respectively; and $F_q$ represents the vertical flux of potential temperature due to processes parameterized by the model (such as the surface fluxes, subgrid-scale turbulent fluxes, gravity wave absorber in the upper part of the domain, etc.). The terms on the right-hand side of (3) are referred to as the convective (or eddy) term, the radiative cooling term, the latent heating term, and the subgrid-scale transport term, respectively.

The interpretation of (3) is that the quasi-equilibrium

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1 Note that the parameterization of microphysics applied in the simulations does not distinguish between latent heating of condensation/evaporation and deposition/sublimation; see Grabowski (1998) for a discussion.
temperature profile is established once the sum of the terms on the right-hand side of (3) becomes zero. For illustration, these terms were estimated for the last 15 days of the PR and IR simulations. The convective transports were directly derived from the vertical velocity and thermodynamic fields. Latent heating was estimated from the water vapor budget. The subgrid-scale transport term was derived as a residual from the quasi-equilibrium balance [i.e., the lhs of (3) assumed zero]. Such profiles for PR and IR simulations are shown in Fig. 11. Profiles for the AR simulation are similar to PR and are not shown.

As shown in Fig. 11, the radiative cooling is balanced mostly by the latent heating. Because radiative cooling is an external forcing for convection, this result is in agreement with other cloud resolving studies (e.g., Sui et al. 1994, Fig. 15; Grabowski et al. 1996a, Fig. 12). The eddy term is small except in the upper troposphere, especially in the simulation PR. This aspect also agrees with previous findings. The large upper-tropospheric eddy term is the only way to balance the strong radiative cooling of the upper troposphere in the PR simulation because the latent heating of a rising saturated parcel is small at very cold temperatures. Note that, analogous to the convective boundary layer, the large eddy temperature flux is consistent with the almost adiabatic upper-tropospheric temperature profile in the PR simulation (and in IR to a smaller degree).

In light of the above discussion, the deviation of the quasi-equilibrium temperature profiles from a typical tropical sounding in all three simulations (cf. Fig. 5) can be attributed to the deviations of the imposed radiative cooling from a typical profile in the Tropics. In the PR simulation, the prescribed radiative cooling is considerably stronger in the upper troposphere (say, above 12 km or 200 hPa) when compared to the observational estimates (e.g., Yanai et al. 1973; Cox and Griffith 1979, among others). The additional cooling in the upper troposphere lowers the upper-tropospheric mean temperature, as predicted by (3). In the IR simulation, on the other hand, the situation is more complicated because of the interaction of radiative fluxes with water vapor and with clouds. However, without doubt the average radiative cooling profile in the IR simulation (Fig. 2b) is not representative of the observed tropical profile either. The upper-tropospheric cooling is too weak, likely because of the extreme (perhaps unrealistic) dryness of the mid- and upper troposphere outside the area featuring convection and clouds.
b. The mean large-scale circulation

In order to investigate the role of the mean temperature profile in establishing the large-scale circulation, let us consider a simple model of the large-scale flow driven by localized convective heating. In this model, the moist convection is represented by a prescribed temperature source, and surface processes are not considered. The model equations for 2D anelastic dry dynamics are

\[
\frac{du}{dt} = -\nabla \pi + kB + D_u, \quad (4a)
\]
\[
\nabla \cdot (\rho_u u) = 0, \quad \text{and} \quad (4b)
\]
\[
\frac{d\theta}{dt} = Q_C + Q_R + D_\theta, \quad (4c)
\]

where \( u = (u, w) \) is the large-scale flow in the horizontal and vertical directions, respectively; \( d/dt = \partial/\partial t + u \cdot \nabla \), \( \nabla = (\partial/\partial x, \partial/\partial z) \); \( \pi \) is the pressure perturbation with respect to a hydrostatically balanced ambient state, normalized by the anelastic reference density \( \rho_o \); \( B \) is the unit vector in the vertical direction; \( \theta = g(\theta - \theta_e)/\theta_w \) is the buoyancy (\( \theta_e \) and \( \theta_w \) are ambient and reference temperature profiles, respectively); and \( g \) is the gravitational acceleration. The potential temperature sources, \( Q_C \) and \( Q_R \), are associated with convective and radiative processes, respectively. The small dissipative terms, \( D_u \) and \( D_\theta \) [in the form of \( K\nabla^2 u \) and \( K\nabla^2 (\theta - \langle \theta \rangle) \)], respectively, are applied to control the amplitude of numerical artifacts. Simulations without dissipation lead to similar solutions as those presented below, albeit with some numerical noise, particularly in the temperature field.

The radiative cooling term \( Q_R \) is prescribed as either 1.5 K day\(^{-1}\) everywhere inside the domain (as in the PR cloud resolving simulation) or equal to the time-averaged radiative temperature tendency in the IR simulation (i.e., the tendency shown in Fig. 2b converted into the potential temperature tendency). The convective heating term \( Q_C \) combines the latent heating, eddy transports, and subgrid-scale transports associated with cumulus convection. It is prescribed in the center of the domain as

\[
Q_C = Af(x)g(z), \quad (5)
\]

where \( A \) is the amplitude, and the functions \( f \) and \( g \) define horizontal and vertical structure of the heating, respectively. The horizontal heating structure is given by

\[
f(x) = \cos\left(\frac{x - x_c}{x_L}\right) \quad \text{for} \quad |x - x_c| \leq x_L \quad \text{and} \quad (6a)
\]

\[
f(x) = 0 \quad \text{for} \quad |x - x_c| > x_L, \quad (6b)
\]

where \( x_c \) is the center of the domain and \( x_L \) is half the horizontal extent of the heating, taken as \( x_L = 1000 \) km for the simulation mimicking the PR cloud resolving simulation and as \( x_L = 500 \) km in the IR case. The vertical heating function is exactly the same as horizontally averaged radiative cooling, that is,

\[
g(z) \sim \langle Q_R \rangle. \quad (7)
\]

The constant \( A \) in (5) assumes that the convective heating has to balance the radiative cooling of the atmosphere when (4c) is considered in a steady-state limit, that is,

\[
\int \rho_o(Q_C + Q_R) dV = 0, \quad (8)
\]

where the integration is over the entire volume of the computational domain.

The dry simulations are conducted with the dry version of the model used in the moist cloud resolving simulations. The model domain is 16 km deep (i.e., no stratosphere) and 4000 km wide. The spatial resolution is 50 km in the horizontal, 0.5 km in the vertical, and the time step is 300 s. The model has periodic lateral boundaries. Boundary conditions at the surface and at the model upper boundary are free-slip rigid lids for the momentum and zero heat flux for the temperature equation.

The model is run for a few weeks using radiative cooling from the PR and IR simulations. Two different initial temperature profiles are used: (i) a typical tropical sounding (taken here as the GATE sounding used to initialize the cloud resolving simulations) and (ii) the quasi-equilibrium temperature profile from the cloud resolving simulations. Because the prescribed convective heating balances radiative cooling at each level, the domain-averaged temperature profiles remain unchanged during these simulations [cf. (3)]. The quasi-equilibrium flows are established within a few days. Figure 12 shows results for the PR radiative cooling; similar conclusions can be drawn for the IR radiative cooling and are not shown. The format of Fig. 12 is the same as Fig. 6 except that twice smaller contour intervals are used. As expected, the simulation applying the quasi-equilibrium temperature profile from the PR simulation (Figs. 12d–f) more closely resembles the PR flow. More importantly, however, the simulation applying cloud resolving quasi-equilibrium temperature profile shows a two-cell vertical structure of the large-scale flow. Consequently, one can conclude that the deviation of the quasi-equilibrium temperature profile from the observed climatology (as represented by the GATE sounding) is indeed responsible for the deviation of the large-scale flow from the predominantly first baroclinic mode structure, as observed in the tropical large-scale circulations.

c. The normal mode analysis of the large-scale circulation

The vertical structure of the large-scale flow can be conveniently analyzed in terms of the vertical normal
modes (e.g., Fulton and Schubert 1985). The normal mode expansion is similar to the Fourier expansion, but it is more natural because it originates from the separation of variables for the primitive equation system. Each normal mode is identified by an index $j$, and it corresponds to a horizontally propagating gravity wave with a phase velocity $c_j$ in the linear dry limit. The normal mode spectra facilitate physical interpretations of model results in terms of the gravity wave dynamics.

The normal mode analysis was performed for the time- and space-averaged virtual potential temperature perturbations and horizontal winds in the vertical domain extending from the surface up to the 25-km-high rigid lid. The temperature perturbations from the ascending branch were averaged in time and space into a single profile. For the horizontal flow, the difference of the right-half domain mean and the left-half domain mean (i.e., a measure of the large-scale convergence into the ascending branch) was used. For the technical details of the normal mode analysis, we refer to Fulton and Schubert (1985). The original Fulton–Schubert formulation has been modified for the anelastic system.

The results for the horizontal winds and temperature for simulations PR, IR, and AR are shown in Fig. 13. The figure shows spectra (defined as the square of the expansion coefficients) as a function of the phase velocity $c_j$. Because the computational domain covers the troposphere and part of the stratosphere, normal modes can be grouped into sets featuring similar structure in the troposphere but different structure in the stratosphere. For instance, in the case of PR horizontal winds (Fig. 13a), the first two modes ($j = 1, 2$; phase speeds 53 and 43 m s$^{-1}$) have similar tropospheric structure. The same is true for the next three modes ($j = 3, 4, 5$; phase speeds 29, 20, and 19 m s$^{-1}$). For the illustration, Fig. 14 shows the tropospheric structure of modes for which the coefficients of the horizontal velocity expansion in the PR case are the largest, that is, modes corresponding to $j = 2, j = 4, j = 7$, as marked in Fig. 13a. These three modes have shapes that are traditionally referred to as the first, second, and third tropospheric baroclinic mode, respectively. As shown in Fig. 13, the large-scale flow predicted by the cloud resolving model is dominated by these three baroclinic modes. The $j = 7$ mode is most dominant in the PR and AR simulations. The IR simulation shows more complex structure in terms of the normal mode decomposition. The normal modes derived in this analysis will be used in the next section to quantify model response in terms of the quasi-two-day oscillations.

5. The quasi-two-day oscillations

Interactions between radiative cooling of the tropical atmosphere and compensating heating associated with deep convection, illustrated in idealized terms in the previous section, give rise to temporal fluctuations of the convection strength, which are referred to as quasi-two-day oscillations (cf. Fig. 1). One can postulate that these fluctuations result from a feedback between in-
tensity of convection and the strength of the large-scale circulation. As illustrated in section 4, balancing the radiative cooling requires an appropriate intensity of the large-scale circulation. A large-scale circulation that is either too strong or too weak would not allow the radiative cooling to be balanced by the large-scale subsidence. Strength of the large-scale circulation, on the other hand, impacts the strength of convection (e.g., through the effect of large-scale low-level convergence). It can then be argued that the two-day-oscillations represent fluctuations of the convection intensity and the large-scale circulation around the condition of perfect balance. However, a closer inspection reveals complexity in the nature of the oscillations.

Figure 15 documents a relation between the strength of convection and the strength of the large-scale circulation. The figure shows evolution of the horizontal wind profile averaged over the left-hand side of the domain, volume-averaged kinetic energy of the large-scale flow (where the large-scale flow at any given time is defined as in Figs. 6, 7, and 8), and the domain-averaged surface precipitation during the last 15 days of the simulation PR. The half-domain horizontal flow illustrates the evolution of the large-scale convergence into the warm SST central part of the domain, with positive (negative) wind contours representing large-scale convergence (divergence) at a given level. The figure illustrates that the periods of enhanced convection during the first week of the 15-day period are indeed associated with the increase of the kinetic energy of the large-scale flow. However, although the strength of convection is clearly in phase with the surface convergence, strength of the flow at higher levels appears to lag in...
time. The reason for this behavior will be clarified through the normal mode analysis (Fig. 17). A temporarily uniform large-scale flow and convection are featured in the second half of the 15-day period.

Figure 16 shows the Hovmöller (x–t) diagrams for the temperature perturbations (defined as deviations from the horizontal mean at a given height) at 2 and 13 km, and the surface precipitation rate for reference, for the last 15 days of the PR simulation. A few features in Fig. 16 are worthy of comment. First, temperature perturbations propagating from the convecting part of the domain toward the descending branch are evident. These perturbations propagate considerably faster than the cloud systems. A good illustration is the movement of temperature anomalies at 2 and 13 km during an active phase of the quasi-two-day oscillation on day 47 as compared to the propagation of the surface precipitation pattern. Figure 16 shows that the perturbations propagate from the center of the domain to the lateral boundaries in about 1–1.5 days, that is, with average speeds of 15–20 m s$^{-1}$. These perturbations also exist in the second half of the period, that is, when the quasi-two-day oscillations are less pronounced. One can argue that the perturbations seem coupled with convection inside the ascending branch during the active phase of the quasi-two-day oscillations. Radiation of internal gravity waves from the convective region follows, in accordance with Mapes and Houze (1995) and Mapes (1998).

To demonstrate the association of the quasi-two-day oscillations with the gravity wave dynamics, an analysis using the normal modes derived in section 4c was performed. In this analysis, the spatially and temporally evolving perturbations of temperature and horizontal velocity fields were decomposed into the vertical normal modes. Figure 17 shows the Hovmöller diagrams of the decomposition coefficients for the $j = 2$, $j = 4$, and $j = 7$ baroclinic modes for the horizontal velocity field. These three modes account for about 65% of the total energy associated with the spectral decomposition, with about equal contributions from all three modes. The next three modes in terms of the energy level are $j = 8$ mode (8%), $j = 1$ mode (5%), and $j = 11$ mode (4%). Figure 17 illustrates the nature of gravity wave interactions between ascending and descending branches. First, as expected, the phase speed of a given mode approximately agrees with the propagation of the amplitude in the physical domain. Second, there are significant differences in the behavior of different modes between the first half of the 15-day period (i.e., when the quasi-two-day oscillations are present) and the second half of that period (i.e., when the oscillations are absent; see Figs. 1a and 15). For instance, the differences for the $j = 2$ mode are significant, whereas the $j = 7$ mode does not seem to be affected by the presence or absence of the quasi-two-day oscillations. This suggests that the quasi-two-day oscillations are not related to the propagation of a single normal mode but are realized by a combination of several modes. The superposition of all the modes results in the fluctuation of the spatially averaged horizontal flow, as illustrated in Fig. 15.

In summary, it is suggested that the quasi-two-day oscillations are associated with the gravity waves radiated from the convective region into the descending branch. According to this picture, the period of these oscillations is dictated by the time required for the waves to propagate around the entire computational domain. Consequently, the period of these oscillations should be proportional to the model domain size. To show that this is indeed the case, an additional 30-day cloud resolving simulation was performed in which the horizontal domain was reduced to half the original size and the spatial scale of the SST variation was reduced in the same proportion. The radiative tendencies were prescribed as in the PR simulation. In general, most features of the PR simulation were well captured in the smaller-domain simulation, including the convection organization and the equilibrium temperature and moisture profiles. However, as illustrated by Fig. 18, the quasi-two-day oscillations were replaced by the quasi-one-day oscillations. This clearly shows that the timescale of the dynamical adjustment associated with these oscillations is indeed related to the size of the computational domain and the speed of gravity waves.

6. Discussion and conclusions

This paper presents results of numerical simulations in which the tropical large-scale circulation is driven by the prescribed spatial distribution of the sea surface temperature (SST). The key point is that spatial resolution is high enough to resolve moist atmospheric convection. The simulations apply an advanced formulation of the surface fluxes (Fairall et al. 1996) and either prescribed radiative temperature tendency profiles (applied homogeneously across the entire domain) or a fully interactive radiation transfer model (Kiehl et al. 1994). With present-day computational resources such simulations can only be considered in 2D, and even then the simulations are computationally demanding; a single 2-month-long simulation discussed in this paper requires about $4 \times 10^5$ time steps and runs for 15–20 days wall clock time on the 64 processor Cray T3D computer.

The two-dimensional idealized model results are not meant to be compared in detail with the natural tropical circulations driven by either zonal or meridional large-scale SST gradient (i.e., the Walker and the Hadley circulations, respectively). One can argue, however, that general features of the simulations should to some extent resemble these natural tropical circulations. For instance, the large-scale flow, featuring outflow from the convecting part in the upper troposphere and the low-level inflow, is consistent with the natural flows and with Raymond (1994). The vertical structure of the circulation, however, is more complex than the first baroclinic mode structure associated with the classic Walk-
Fig. 16. Hovmöller (x-t) diagrams of the potential temperature perturbations from the domain average at a height of (a) 13 and (b) 2 km for the last 15 days of the PR simulation. The temperature perturbations smaller than $-0.5$ K are shown without any shading, and perturbations larger than 0.5 K are shown using dark shading. (c) The surface precipitation intensity (as in Fig. 4a).
Fig. 17. Hovmöller (x–t) diagrams of the expansion coefficients for the horizontal velocity associated with the (a) $j = 2$, (b) $j = 4$, and (c) $j = 7$ baroclinic modes. All three panels apply the same color scale and the blue (yellow) color shows negative (positive) values of the expansion coefficients. The phase speed of a given mode is shown as a solid line in the upper part of each panel.
FIG. 18. (top) Evolutions of the sources of the domain-averaged density-weighted temperature and (bottom) the precipitable water in a simulation with the horizontal domain of 2000 km, that is, one-half of the domain used in the PR, IR, and AR simulations. The experiment is carried out for only 30 days, which is sufficient to reach the quasi-equilibrium state. The oscillations in the second half of the experiment have a period of about 1 day, that is, about one-half of the period observed in PR, IR, and AR. Compare the top with the bottom of Fig. 1.

The simulations with prescribed radiative temperature tendencies (i.e., the simulation reported in G98 and the PR simulation of this paper) evolved into a regime featuring weakly coupled circulations in the lower and upper halves of the troposphere, weak convection in the center of the domain, and unrealistically low CAPE (around 100 J kg$^{-1}$). The IR temperature profiles featured a more stable lower half of the troposphere and a less stable upper half when compared to a typical tropical sounding (Fig. 5b). These dramatic differences in the large-scale circulations reflect the differences in the temperature tendencies due to radiative processes in both PR and IR simulations. The additional AR experiment, having interactive radiation but an adjusted horizontally averaged radiative tendency to match the PR tendency profile, demonstrated that the strength of convection was related to the mean radiative tendency profile. However, the area covered by convection seemed determined by the difference between radiative transfer in cloudy and clear-sky parts of the domain. The net energy loss due to radiative processes averaged over the entire domain was approximately the same in all three experiments.

Indirectly, numerical simulations presented in this paper demonstrate that a simple picture in which the tropical atmosphere experiences a net cooling (approximately homogeneous in both horizontal and vertical) by radiative processes is an overidealization. Inclusion of the radiation transfer model in the AR and IR simulations resulted in a dramatic modification of the large-scale flow compared to the PR simulation. Such a statement is consistent with recent results of Raymond (2000), who goes a step further and claims that the interaction of radiation with spatially and temporarily varying water vapor and cloud condensate might be a missing link in the large-scale atmospheric dynamics in the Tropics.

In terms of the water and energy budgets, the two-day oscillations observed in the simulations reported herein are reminiscent of the quasi-two-day disturbances observed in the Tropics (e.g., Takayabu et al. 1996; Chen and Houze 1997; Haertel and Johnson 1998). For instance, in both the model simulations and in the observations, the active phase of the oscillation is associated with enhanced convection (as measured by the surface precipitation) and enhanced surface fluxes. The explanations for the two-day disturbances proposed by Takayabu et al. (1996) and by Chen and Houze (1997) both refer to the interaction of deep convection with equatorially trapped waves. The quasi-two-day oscillations present in our numerical simulations, on the other hand, are a consequence of the interaction of deep convection with convectively generated gravity waves in a periodic computational domain. In particular, the period of the oscillations is directly affected by horizontal extent of the computational domain. To what extent the resemblance between the tropical quasi-two-day disturbances and the two-day oscillations discussed in this paper is coincidental remains to be seen.

Numerical experiments reported in this paper are just a first step in our long-term goal to understand the in-

2000–KM DOMAIN RUN

sources of temperature (K day$^{-1}$)

condensation

surface flux

radiation

sources of precipitable water (kg m$^{-2}$ day$^{-1}$)

evaporation

precipitation

time (days)

FIG. 18. (top) Evolutions of the sources of the domain-averaged density-weighted temperature and (bottom) the precipitable water in a simulation with the horizontal domain of 2000 km, that is, one-half of the domain used in the PR, IR, and AR simulations. The experiment is carried out for only 30 days, which is sufficient to reach the quasi-equilibrium state. The oscillations in the second half of the experiment have a period of about 1 day, that is, about one-half of the period observed in PR, IR, and AR. Compare the top with the bottom of Fig. 1.
teractions between resolved cumulus convection and the large-scale circulations in tropical dynamics. As discussed in the introduction, the incorporation into a single, dynamically consistent framework of cloud dynamics, microphysics, and large-scale flow is the hallmark of our approach.

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