Explicit Simulations of the Intertropical Convergence Zone

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ABSTRACT

The intertropical convergence zone (ITCZ) is one of the most important components of the global circulation. In order to understand the dynamical processes that regulate its formation, latitudinal preference, and structure, explicit two-dimensional numerical modeling of convection on an equatorial beta plane was conducted with a nonhydrostatic cloud-system-resolving model. The model was forced by energy fluxes associated with constant sea surface temperature (SST) and by horizontally homogeneous radiative cooling.

Two distinct patterns were identified for the spatial distribution of convective activity in the Tropics. The first was characteristic of enhanced off-equator convection, namely, a double ITCZ-like morphology (one more salient than the other) straddling the equator during the early period of the integration. The second featured enhanced equatorial convection, namely, a single ITCZ-like morphology on the equator during the later quasi-equilibrium period. Diagnostic analysis and two additional experiments, one excluding surface friction and the other having time- and space-independent surface fluxes, revealed that the wind-induced surface flux variability played an essential role in the development and maintenance of the equatorial maximum convection. Surface friction was largely responsible for the early asymmetric convective distribution with respect to the equator in the control simulation and acted to flatten the convective peaks.

One important discrepancy from observations concerned the too-weak trade wind convergence around enhanced convective regions. This unrealistic feature suggested that, as well as the meridional dynamics, latitudinal SST gradients, large-scale forcing, and other physical processes regulate the observed ITCZs.

1. Introduction

The intertropical convergence zone (ITCZ) refers to the narrow and approximately east–west-oriented belt of concentrated vigorous cumulonimbus convection in the Tropics and represents the ascending branch of the meridional Hadley circulation. Observations show that the ITCZ typically resides between the latitudes 4° to 12° away from the equator over warm oceanic regions (e.g., Waliser and Somerville 1994). However, the precise latitudinal location varies greatly with season and longitude. Moreover, a pair of ITCZs straddling the equator also occurs (e.g., Lietzke et al. 2001; Zhang 2001; Halpern and Hung 2001; Liu and Xie 2002), as well as the common situation where a single ITCZ is located at or away from the equator.

The physical mechanisms regulating the formation and latitudinal preference of the ITCZ have been a subject of numerous observational, theoretical and numerical modeling investigations. The earliest attempts tried to relate the spatial distributions of sea surface temperature (SST) to the spatial structure of tropical convection, motivated by the observed high correlation between convective enhancement and warm SST forcing (Bjerknes et al. 1969; Graham and Barnett 1987). This hypothesis is supported by a number of numerical simulations (e.g., Pike 1971; Manabe et al. 1974; Goswami et al. 1984).

Although it has a strong influence on observed tropical convection and ITCZs, SST forcing alone cannot explain all observed features, and considerable variation exists in the relationship of convection to the underlying SST distribution. For instance, many observational studies showed that the highest SST is often not collocated with the ITCZ (e.g., Ramage 1974; Sadler 1975; Hastenrath and Lamb 1977; Lietzke et al. 2001). In addition, general circulation modeling (e.g., Hayashi and Sumi 1986; Hess et al. 1993; Lietzke et al. 2001) showed that double ITCZs develop straddling the equator even if the SST maximum is at the equator. Furthermore, a well-defined ITCZ can still occur in numerical simulations with globally uniform SST (e.g., Sumi 1992; Chao 2000; Kirtman and Schneider 2000), implying that inhomogeneous SST forcing might not be necessary. From these observational facts and modeling studies, it is concluded that dynamical processes likely play an important role in regulating the observed ITCZs.

Charney (1971) put forward an explanation for the
ITCZ formation in terms of two competing processes, namely Ekman pumping and moisture availability. The former is proportional to the Coriolis parameter and thus increases poleward, whereas the latter generally increases equatorward. Therefore, their combined effect results in the moisture supply peaked at a finite distance from the equator, accounting for an off-equator ITCZ.

Several hypotheses for the off-equator ITCZ were proposed based upon low-level convergences. Holton et al. (1971) argued that the ITCZ is favored at the latitudes where the frequency of zonally propagating disturbances equals the Coriolis frequency, and maximum boundary layer convergence occurs. This theory was furthered by Chang (1973) who showed that the mechanism is indeed responsible for the development of the ITCZ in a numerical model. In contrast, observational studies found that the ITCZ does not owe its existence to zonally propagating synoptic-scale disturbances, in the sense that it would still exist in their absence (Gu and Zhang 2002). Lindzen (1974) found that there exists a longitude-independent wave–CISK (conditional instability of the second kind) mode with a period of about 5 days. This oscillation, when coupled with the traveling waves of the same period generated by the inhomogeneous surface conditions, leads to the maximum low-level convergence several degrees away from the equator. Based upon linear shallow-water theory, Waliser and Somerville (1994) also demonstrated that the largest local low-level convergence is produced when a heat source is located at a finite distance from the equator.

Recent studies showed that the character of tropical convection is affected by the cross-equatorial pressure gradients and radiative–convective instability. Tomas and Webster (1997) noted that, in regions of a substantial cross-equator surface pressure gradient, a local anticyclonic circulation exists on the low pressure side of the equator, and the flow meets the criterion for inertial instability. It was hypothesized that the atmospheric response to the absolute vorticity distribution results in a low-level convergence–divergence doublet, which is important in determining the strength and location of convection. On the other hand, Raymond (2000) examined the radiative role of convective clouds in the dynamics of the Hadley circulation. The modeling results demonstrated that cloud–radiation interaction is strong enough to drive a circulation that is comparable in intensity to the observed global mean Hadley circulation.

Studies of the ITCZ are further complicated by the sensitivity of the modeled ITCZ location and structure in GCMs to convective parameterizations. For example, in aquaplanet modeling with globally uniform SST, Chao (2000) and Chao and Chen (2001) found that a double ITCZ can evolve into a single one by simply switching the relaxed Arakawa–Schubert parameterization (Moorthi and Suarez 1992) to the moist convective adjustment scheme (Manabe et al. 1965). The high sensitivity of the ITCZ behavior to convective parameterization was also reported in aquaplanet models with an equatorial SST maximum (Hess et al. 1993; Numaguti 1993).

The uncertainty resulting from convective parameterization is a serious disadvantage in using GCMs to investigate the physical processes governing the observed ITCZs. In this study, we adopt an explicit approach—cloud-system-resolving modeling (CSRM)—that avoids the use of any convective parameterization. Presently, computer capability limits our simulations to two spatial dimensions; namely, a latitude–height cross section across the equator on an equatorial beta plane. We have two major objectives. First, we examine patterns of convective activity generated solely under the influence of the earth’s rotation and, in particular, whether concentrated tropical convection (i.e., the ITCZ-like features) resides either at or off the equator. Second, we examine the mechanisms responsible for the simulated ITCZ-like concentrated tropical convection.

The numerical model and design of the numerical experiment are described in the next section. In section 3, the simulation results are detailed. Section 4 is devoted to the physical interpretation of the simulated convective features through diagnostic analyses and two sensitivity experiments. The paper concludes in section 5.

2. Numerical model and experimental design

We use the two-dimensional Eulerian version of the nonhydrostatic Eulerian/semi-Lagrangian anelastic model (Smolarkiewicz and Margolin 1997). The 16 000 km × 24 km (north–south oriented) computational domain represents an equatorial beta plane bounded with rigid walls and centered on the equator. The horizontal and vertical grid spacing are 5 km and 0.3 km, respectively. Note that the 5-km grid length is a practical compromise made to enable longtime, large-domain simulations. It allows the mesoscale organization of convection to be treated explicitly, but not individual convective cores. This grid size has been used in previous studies of tropical convection (e.g., Held et al. 1993). Rigid and free-slip vertical boundary conditions are employed at the top and bottom of the domain. At the lateral boundaries, the potential temperature and water vapor mixing ratio equal their respective domain-averaged values at every time step. Otherwise, undesired large temperature and water vapor gradients could occur near the lateral boundaries when the modeled atmosphere departs sufficiently from the initial state. A 500-km-wide and a 6-km-deep absorbing layer at the lateral and top vertical boundaries, respectively, damp propagating gravity waves that could otherwise be reflected unrealistically.

The bulk cloud microphysical parameterization uses a two-category warm rain scheme (Grabowski and Smolarkiewicz 1996) and a two-category ice scheme (Grabowski 1999). The domain lies over an ocean having a constant SST (302.5 K). The surface moisture and sen-
sible heat fluxes are calculated using a simplified version of the Tropical Ocean Global Atmosphere Coupled Ocean–Atmosphere Response Experiment (TOGA COARE) surface flux algorithm (Fairall et al. 1996). Rather than using cloud-interactive radiation, a time-independent and horizontally uniform radiative cooling (a constant value of $-1.5 \text{ K day}^{-1}$ below 12 km decreasing linearly to zero at the model top) is specified. This simplification facilitates the interpretation of the simulation results and exploration of the fundamental physics of ITCZs.

The simulation starts from a resting atmosphere. The initial temperature and moisture field are based upon the averaged condition over the Intensive Flux Array (IFA) during the 19–26 December 1992 period of TOGA COARE. Small random perturbations of the potential temperature and water vapor mixing ratio fields initiate convection.

3. Description of simulation

Our 100-day simulation enables the simulated atmosphere to attain a state of statistical quasi-equilibrium. Because the trade wind is not prescribed, the modeled ITCZ-like circulation must result solely from the atmospheric response to tropical convection. As indicated in the Hovmöller diagram of surface precipitation (Fig. 1), the convective pattern in the Tropics during the early period of integration is dramatically different from the late quasi-equilibrium stage, corresponding to off-equator and equatorial ITCZ-like features, respectively. These two distinct phases are now described.

a. Off-equator maximum convection stage

The off-equator ITCZ-like pattern roughly corresponds to the first 40 days of integration. Characteristically, the convective activity displays a dual maximum straddling the equator, but is not symmetric across the equator: one is more salient than the other. The space–time surface precipitation distribution in Fig. 1 illustrates the latitude-dependent behavior of convective activity. First, convection at high latitudes is frequent, but randomly distributed and short-lived. In contrast, tropical convection is comparatively strong, persistent, organized and aggregated. Second, vigorous convection in the Tropics is preferentially off-equator (at least several hundred kilometers away from the equator): it rarely occurs near the equator. Third, convection is asymmetric about the equator and preferentially develops in the Northern Hemisphere. Finally, once vigorous convection occurs on one side of the equator, convective activity is commonly suppressed on the other side. Another noticeable feature is the large clear region surrounding the tropical convective systems.

The precipitation rate distribution averaged over the first 40 days (Fig. 2a) shows that the enhanced convective region is concentrated at approximately 1200 km from the equator in the Northern Hemisphere, and the secondary peak is located roughly 500 km from the equator in the Southern Hemisphere. In contrast, precipitation in the neighborhood of the equator is suppressed significantly, consistent with the temporal and spatial rainfall distribution. Outside the tropical area, the precipitation intensity does not show marked spatial dependence, except in the outermost 500-km-wide sponge layers, where convection is suppressed completely. Figure 3a further demonstrates the meridional convective variability in terms of the spatial distribution of cloud fraction. In calculating the cloud amount, 100% cloudiness is assumed over a grid box when the total condensate (i.e., the sum of cloud water, rain water, and cloud ice) exceeds 0.01 g kg$^{-1}$. The fractional cloudiness clearly displays an off-equator maximum with a value exceeding 25% about 1500 km from the equator in the Northern Hemisphere, a secondary extremum in the Southern Hemisphere, and an equatorial minimum. This spatial distribution is well-correlated with the precipitation variability in Fig. 2a.

Figure 4 shows the perturbation fields averaged over a period of 10 days (from days 16 to 25) based on the hourly archived dataset. The fields are smoothed with a 500-km running mean filter to eliminate the smaller-scale noise and highlight the organized features. The tropical meridional circulation in Fig. 4a displays a striking asymmetry with respect to the equator. In the Northern Hemisphere, equatorward flow prevails at upper levels and poleward flow at lower levels, leading to upper-level convergence and low-level divergence at the equator. The opposite circulation pattern occurs at the poleward flank of the concentrated convection about 1200 km from the equator. The flow structure supports a deep off-equator ascent and a wide equatorial descent (see the shading) and is consistent with weak equatorial convection and active off-equator convection. A similar, albeit much weaker, meridional circulation occurs in the Southern Hemisphere and is evidently related to the weak convective maximum there. It should be pointed out that although a Hadley-like circulation develops, the meridional flow at both low and upper levels is rather weak and localized, in contrast with some previous modeling results (e.g., Waliser and Somerville 1994; Raymond 2000) using cumulus parameterizations. However, a strict comparison is meaningless because our model has neither latitudinal SST variation nor cloud–radiation interaction, unlike the GCMs. Another noticeable feature is the equatorial southerly flow at 7–11 km and northerly flow at 4–6 km. The existence of midlevel divergence around the equator illustrates that the subsiding warm air partially moves away from the equator before reaching the lower troposphere. The similar double-celled meridional circulation behavior has been documented in the zonally symmetric tropical atmosphere from observations (e.g., Johnson et al. 1999; Mapes 2001). Although instantaneous perturbations are sub-
Fig. 1. Space–time distributions of surface precipitation rate during (a) days 1–25, (b) days 26–50, (c) days 51–75, and (d) days 76–100. The light and dark shading correspond to rainfall intensity greater than 1 and 10 mm h$^{-1}$, respectively. The equator is located at the center of the domain.
The most striking feature in Fig. 4b is the deep equatorial easterly wind prevalent throughout the troposphere, peaking just below the tropopause. Characteristically, outside the Tropics the wind perturbation is easterly at low levels and mostly westerly at upper levels. The two subtropical westerly jets in the upper troposphere are intimately associated with the effects of the Coriolis torque on the poleward flow of the meridional circulation. This is consistent with the conceptual explanation that the westerly jets originate from the angular momentum transports of Hadley cells (e.g., Held and Hou 1980; Lindzen and Hou 1988). In comparison with the counterparts in the real atmosphere; however, the westerly jets are at the wrong latitudes and much too weak because the observed subtropical jets are influenced strongly by both stationary and transient eddy flux convergences, which are absent in our two-dimensional model setup.

The temperature field in Fig. 4c is dominated by widespread cooling relative to the initial state. The warmest air resides over the equator, resulting from adiabatic subsidence, and the coldest occurs about 6500 km from the equator in each hemisphere, maintaining a weak equator-to-pole temperature gradient. (The reversed temperature gradient near the boundaries is artificial, the result of the lateral boundary conditions and damping zones.) Figure 4d features weak moist perturbations in the planetary boundary layer and strong dry perturbations in the free troposphere. The equatorial region and the adjacent tropical atmosphere undergo prominent drying due to compensating subsidence, which is a response to off-equator vigorous convection.

The pressure distribution in Fig. 4e hydrostatically corresponds to the temperature perturbation and exhibits a relatively low pressure in the lower troposphere and a relatively high pressure aloft in the Tropics. The lowest surface pressure occurs in the concentrated convection...
region, whereas the surface highs are positioned about 6500 km away from the equator in accordance with the respective warmest and coldest locations. The pressure gradient is mainly pole-to-equator at lower layers, reversing at upper levels.

**b. Equatorial maximum convection stage**

After the simulation achieves a quasi-equilibrium state, the aforementioned off-equator ITCZ-like pattern transforms into a single equatorial ITCZ-like morphology as evinced in the space–time precipitation distribution (Fig. 1). Long-lasting convection, or convection aggregated on mesoscales, is concentrated in a narrow area around the equator, whereas flanking active convection is scarce. Convective activity is not continuous near the equator. Instead, groups of convective clusters are intermittent with a regular period of about 2 days. Approximately, each episode persists for 1 day, followed by a 1-day suppression. The periodic enhancement and suppression of convective activity are a result of the destruction and recovery of convective available potential energy (CAPE). This is clearly demonstrated by the out-of-phase correlation of precipitation and CAPE spatially averaged over a 1500-km-wide region around the equator in Fig. 5. Once deep convection breaks out, the atmosphere is stabilized via latent heating in the middle and upper troposphere and downdraft-induced evaporative cooling in the boundary layer: convection cannot last forever over a limited area. During a suppressed episode, the persistent moisture and sensible heat transports from the underlying warm ocean, as well as the imposed radiative cooling and horizontal moisture convergences in the lower troposphere, restore CAPE gradually. On average, the CAPE peaks approximately 15 h earlier than the precipitation.

Figure 2b presents the spatial distribution of the surface precipitation rate averaged over the last 60 days (from days 40 to 100). Enhanced convection extends approximately 1000 km across the equator, and the surrounding convective activity is substantially reduced in the tropical region: a precise reversal of the earlier convective activity. The strong equatorial convection is also reflected in cloudiness, which has a salient maximum of over 35% at the equator, and a minimum of less than 10% in the nearby latitudes (Fig. 3b).

Figure 6 illustrates the atmospheric structure during the quasi-equilibrium stage. The tropical meridional wind in Fig. 6a displays a wavelike vertical structure of wavelength half of the tropospheric depth and is opposite in sign in the two hemispheres. The attendant convergence–divergence and ascent–descent distribution are exactly opposite to those in the off-equator active convection phase. They are consistent with enhanced equatorial convective activity that draws in warm and moist boundary layer air from both sides of the equator, transports it upward, and detrains it near
the tropopause. We speculate that the weak divergence around the 5-km level is due to the shallower convection (cumulus congestus) with tops near the melting level (Johnson et al. 1999; Liu et al. 2001a,b), while the overlying weak convergence is attributable to the adjacent descending warm and dry air that could not reach the lower troposphere and diverges in the middle troposphere. Alternatively, this two-cell vertical structure was interpreted as a consequence of the vertically varied static stability in the tropical troposphere by Mapes (2001). In high latitudes, the mean meridional wind is weak and does not possess coherent large-scale circulations as in the off-equator convective maximum. Again, the meridional wind is weak and localized, suggesting that a moderate SST gradient and/or other physical processes, such as interactive radiation and large-scale forcing, are necessary to generate realistic trade winds and Hadley circulations.

The zonal wind in Fig. 6b is much stronger than the meridional component. Interestingly, the equatorial easterly wind and flanking westerlies prevail regardless of whether enhanced convection occurs at or off the equator. Unlike the earlier off-equator maximum convection episode, however, the equatorial easterly is uniformly distributed in the vertical due to convective mixing and does not possess a jetlike structure. The high latitudes are dominated by easterlies in the lower troposphere except in the vicinity of the lateral boundaries, similar to the enhanced off-equator convection stage.

The temperature perturbation in Fig. 6c is characterized by a warming relative to the initial atmosphere in the lower tropical troposphere and cooling elsewhere. The prominent two warm centers are located at about 1000 km from the equator, obviously a result of the subsidence warming surrounding the equatorial convection. The cold center is situated at about 1500 km from the lateral boundary, an artifact of the lateral boundary conditions as indicated earlier.

An equatorial easterly flow occurs irrespective of the latitudinal position of the convective peak (see Figs. 4b and 6b). Figure 7a shows the evolution of this tropospheric easterly wind in terms of the zonal-average over a 1000-km domain centered at the equator. Several characteristics are salient. First, the development of the equatorial easterly wind starts in the upper troposphere, gradually expands downward and reaches the surface at about 25 days into the simulation. Second, an easterly jet is maintained in the upper troposphere during the first 40 days. Third, vertically uniform easterlies prevail and there is little shear throughout the troposphere during the enhanced equatorial convection.

In order to quantify the physical processes in the development of the equatorial easterlies, the zonal momentum budget is investigated. In a two-dimensional (latitude–height) anelastic framework, the zonal momentum equation is given as

\[ \frac{\partial \bar{u}}{\partial t} = -\frac{1}{\rho} \frac{\partial \bar{p} \bar{u}}{\partial y} - \frac{1}{\rho} \frac{\partial \bar{p} \bar{w}}{\partial z} + f \bar{v} + \bar{D}_u, \]  

(1)
where the notations are standard, and the overbar represents the spatial average.

Figure 8 displays the zonal wind tendency due to horizontal momentum flux convergence \((-\nabla\cdot(p\mathbf{u}))\), vertical momentum flux convergence \((-\nabla\cdot(p\mathbf{w}))\), and Coriolis torque \(\mathbf{f}_u\), from days 10 to 60. In the off-equator concentrated convection stage, the horizontal convergence term (Fig. 8a) is the major source for the upper-level easterlies. However, the upper-level easterly momentum originates from the equatorward meridional flow outside the budget region through the Coriolis force as inferred from Fig. 4a. The vertical convergence term (Fig. 8b) accounts for the easterly wind generation in the middle and lower troposphere. Evidently, the gradual downward extension of the easterly momentum source during the early 25 days is consistent with the persistent downward expansion of easterly flow as evinced in Fig. 7a. After the maximum convection is displaced to the equator (roughly 40 days into the integration), the two branches of equatorward meridional flow associated with the double-celled vertical circulation in Fig. 6a produce easterly momentum by the Coriolis torque. The easterly momentum subsequently converges near the equator and gives rise to two easterly momentum sources located near the surface and at 7–10 km (Fig. 8a), respectively. The equatorial convection transports the easterly momentum upward, maintaining the easterly flow in the 2–6-km layer and above 11 km (Fig. 8b).

Although the Coriolis force is essential to the easterly momentum generation outside the budget domain (not shown), its contribution within the budget domain is of secondary importance because of the small Coriolis parameter (Fig. 8c). As expected, the Coriolis torque produces easterly momentum in the equatorward meridional flow and westerly momentum in the poleward flow. The subgrid diffusion \(D_u\) is weak except in the planetary boundary layer where it acts to reduce the easterly wind speed (not shown).

The foregoing analysis suggests the following physical basis for the formation and sustenance of the equatorial easterlies. During the early stage, active off-equator convection transports low-level mass upward and deposits it in the upper troposphere. The detrained mass moves toward the equator from both hemispheres, as evinced by the meridional wind in Fig. 4a, and generates an equatorial easterly wind at upper levels due to the turning action of the Coriolis force. The meridional flow
converges over the equator, causing equatorial subsidence. This slow descent transports the easterly momentum and gradually extends the easterly wind into lower layers and eventually down to the surface. During the quasi-equilibrium stage, convective activity is concentrated at the equator, and the meridional circulation is reversed. In contrast to the early period, easterly momentum is generated in the lower troposphere and also within an elevated convergence layer, converges toward the equator and is subsequently transported vertically by attendant convection. Because of the strong, fast, and efficient convective transport, the well-mixed easterly wind is striking during the equatorial ITCZ episode. It should be pointed out that the above explanation may be only applied to our two-dimensional framework. In the real world, the transient meridional circulation associated with the seasonal cycle is crucial for maintaining the climatological easterly wind in the equatorial upper troposphere (Lee 1999).

4. Physical interpretation

As reviewed in the introduction, quite a few hypotheses have been proposed for the observed off-equator ITCZs. However, most arguments do not apply to our simulated off-equator maximum convection in a two-dimensional equatorial beta plane with uniform SST. Moreover, none of the previous theoretical arguments are relevant to the single equatorial convective peak occurring in the quasi-equilibrium stage. In the following, thermodynamical and dynamical arguments are presented for the equatorial and the off-equator ITCZ morphologies.

a. Surface friction

Surface friction influences ITCZs and attendant convection in various ways. First, the friction-induced Ekman pumping, which is a function of the Coriolis parameter, preferentially supports high-latitude convection (Charney 1971). Second, according to Chao and Chen (2001), surface friction increases the surface energy intake of the boundary layer air converging toward convective areas because of a longer inward-spiraling path than in the frictionless situation, which also favors higher-latitude convection. Finally, surface friction decreases the wind speed near the ground, affecting the wind-induced surface flux variability, which is a mechanism responsible for the simulated strong equatorial convection (see later discussion).

To quantify the potential role of surface friction in the simulated convective behavior, we performed a 100-day sensitivity experiment, which is identical to the control run except that the surface friction is excluded. As expected, absence of the friction enhances the wind speed near the surface. Increased moisture and sensible heat transports from the underlying warm ocean lead to an overall stronger convective activity compared to the control experiment. Nevertheless, the integration again has two distinct phases, namely off-equator and equatorial concentrated convection in terms of the Hovmöller diagram of surface precipitation rate (not shown). The instantaneous convective behavior and the spatial dependence are barely affected although the off-equator concentrated convection only lasts for about 1 month, roughly 10 days shorter than in the control simulation.

The close similarity between the two experiments is further demonstrated by comparing the time-averaged spatial distributions of precipitation rate in Fig. 9 against those in Fig. 2. The convective peak is more pronounced in the frictionless simulation. Another apparent feature is that the early double convective peaks straddling the equator, although less persistent, are comparable in intensity unlike their counterparts in the control simulation with one much stronger than the other. When surface friction is absent, the low-level easterly momentum in the converging inflow toward the equatorial region is amplified. This intensified momentum generation results in a stronger tropical easterly flow as indicated in Fig. 7b. The intensified surface wind leads to large surface energy fluxes and thus favors convective development in the equatorial region. This is consistent with the aforementioned swifter transition from off-equator to equatorial regime compared to the control simulation. Note that the equatorial east wind tends to gradually intensify with time and does not realize a steady state during the 100 days of integration.
The preceding discussion reveals that the water vapor and temperature distributions contradict the surface flux distributions. In other words, an inhomogeneous surface wind may be important in regulating the energy transports from the underlying warm ocean, as demonstrated in space–time distributions of surface wind speed (Fig. 11) and evolution of the spatially averaged surface flux and wind speed over a 1500-km equatorial area (Fig. 12). Evidently, a positive correlation exists between surface wind and heat flux intensity; in general, large wind speeds are accompanied with large surface fluxes. A swift transition is noticeable around 25 days into the simulation, as well as the variations on short time scales associated with convective modulations. Before the easterly wind gets firmly established near the surface, the wind speed maximum is positioned at more than 1000 km from the equator in the Northern Hemisphere, correspondent with the surface flux maximum in the concentrated convective area, whereas the weak equatorial wind is well correlated with the surface flux minimum there. Quantitatively, the two individual wind components (not shown) are comparable except in the Tropics, and both exhibit spatial patterns similar to the total wind speed. The meridional wind accounts for most of the total wind maximum, but the weak zonal wind near the equator causes the significant minimum in the total wind speed. For the most part, we speculate that the weak equatorial zonal wind is due to the small Coriolis parameter and accounts for the weakened equatorial surface fluxes, which is undesirable for equatorial convection and indirectly beneficial to the enhancement of off-equator convection.

During the period of equatorial concentrated convection, the wind speed peaks at the equator and decreases rapidly toward high latitudes. Apparently, the enlarged surface fluxes around the equator occur after the easterlies reach the surface. On average, the peak wind speed is roughly 150% as large as the value outside the Tropics. It is the strong equatorial wind speed that is responsible for the surface flux maximum therein. This implies that the development of the enhanced equatorial convection results from wind-enhanced equatorial surface fluxes.

c. Uniform surface fluxes

The foregoing analyses suggest that wind-induced surface flux variability is a viable mechanism for the development of the simulated ITCZ-like features. During the early period of simulation, the weak Coriolis acceleration generates the weak zonal flow around the equator. This leads to an equatorial surface flux minimum which, in turn, subdues the equatorial convection. As demonstrated in Fig. 7a, an equatorial easterly wind gets gradually established in the upper troposphere and spreads downward during the off-equator ITCZ period. When the equatorial easterly reaches the surface, the equatorial surface flux is increased and eventually ex-
Fig. 10. Space–time distributions of surface fluxes during (a) days 1–25, (b) days 26–50, (c) days 51–75, and (d) days 76–100. The light, moderately dark, and heavy dark shadings correspond to values greater than 100, 140, and 180 W m$^{-2}$, respectively.
Fig. 11. Space–time distributions of surface wind speed during (a) days 1–25, (b) days 26–50, (c) days 51–75, and (d) days 76–100. The shading scale is shown at the upper-right corner (in m s\(^{-1}\)).
circulation. The physical processes governing its main-

5. Concluding discussion

The ITCZ is a key feature of the tropical large-scale
circulation. The physical processes governing its main-
tenance, latitudinal position and structure are still poorly
understood. Understanding the responsible mechanisms
is further complicated by the strong dependence of
ITCZs in GCMs on cumulus parameterizations. Also,
GCMs contain complex interactions that are difficult to
unambiguously quantify. To simplify the physics and
overcome this vexing uncertainty, we adopted an ex-

The persistence of suppressed equatorial convection
illustrates that the wind-induced surface fluxes around the
equator are substantially enhanced due to the equatorial
wind intensification. This wind-induced surface flux enhancement supports the
equatorial convective activity, eventually leading to the
development of a single ITCZ-like feature at the equator.
The role played by the wind-induced surface flux varia-
tions in controlling the tropical convection is con-
firmed by an additional experiment in which the surface
fluxes are uniformly prescribed. In particular, the off-
equator concentration of convective activity is a robust
feature in this sensitivity experiment, suggesting that
the wind-induced surface flux enhancement is crucial
to the formation of an equatorial ITCZ-like feature.

The off-equator ITCZ-like feature occurs in all of the
Fig. 13. Same as in Fig. 1, but for the experiment with uniform surface fluxes.
three simulations, but none of the existing hypotheses provide an adequate interpretation. Recent work by Liu and Moncrieff (2002) suggests that the simulated off-equator peak convection results from two competing dynamical mechanisms associated with the latitudinal variation of the Coriolis parameter. The first is rotation-induced trapping and accumulation of compensating subsidence warming and drying in response to convective heating: this mechanism favors an equatorial ITCZ. The second concerns the rotation-induced strengthening of the low-level convergence in the heating area, which prefers an ITCZ displaced from the equator. These two conflicting physical processes compromise to displace the maximum convection a finite distance from the equator. Although this theoretical argument is attractive, further studies are needed to substantiate its relevance to the simulated off-equator ITCZ-like pattern herein.

Because of our highly idealized experimental setup, it is not surprising that the model results differ from or even contradict some observed facts. For example, a convective minimum near the equator is often observed in the western Pacific warm pool where quite uniform SST exists, distinct to the equatorial convective maximum in the quasi-equilibrium stage. Another noticeable discrepancy from observations concerns the magnitude of shallow warm clouds; the model warm clouds account for roughly 14% of total rainfall, less than the 20% maximum in the quasi-equilibrium stage. Another noticeable discrepancy from observations concerns the magnitude of shallow warm clouds; the model warm clouds account for roughly 14% of total rainfall, less than the 20%

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