A Simple Moist Tropical Atmosphere Model: The Role of Cloud Radiative Forcing

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ABSTRACT

A simple moist model for the large-scale tropical atmospheric circulation is constructed by combining the simple models of Gill and Neelin and Held. The model describes the first baroclinic mode of the moist troposphere with variable ''gross moist stability'' in response to given thermodynamic forcing from surface evaporation and atmospheric cloud radiative forcing (CRF), which is a measure of the radiative effects of clouds in the atmospheric radiative heating. When the present model is forced solely by the observed atmospheric CRF, quantitatively reasonable Hadley and Walker circulations are obtained, such as the trades, the ascending branches in the intertropical convergence zone (ITCZ) and the South Pacific Convergence Zone (SPCZ), as well as the descending branches in the cold tongue and subtropics. However, when the model is forced only by the observed surface evaporation, the Walker circulation totally disappears, and the Hadley circulation reverses. These results indicate that, in the context of a moist dynamic model, the spatial variations of atmospheric CRF are more important in terms of driving and maintaining the Hadley and Walker circulations than the spatial variation of surface evaporation.

1. Introduction

A successful conceptual model for the tropical large-scale atmospheric circulation is that of Gill (1980) which proposed that the circulation is a linear response to atmospheric diabatic heating. Latent heating and atmospheric cloud radiative forcing (CRF) are the two major sources of diabatic heating for the tropical atmosphere (Ramanathan 1987; Webster 1994). Atmospheric CRF is the difference between the radiative heating of the atmosphere for average cloudiness and that of clear skies. Thus, atmospheric CRF is a meaure of how clouds alter the radiative heating of the atmosphere (Tian and Ramanathan 2002). An important feature of these two sources of diabatic heating is that they reinforce each other (Webster 1994). That is, the large-scale spatial patterns of atmospheric CRF and latent heating, as well as their anomalies, are similar. Furthermore, seasonal variations of those two sources of heating are, for the most part, in phase (Sohn 1999; Bergman and Hendon 2000b). The differential gradients of atmospheric CRF across the Pacific are about one-fourth to one-third of those of latent heating (Webster 1994). Thus, according to the linear Gill model, atmospheric CRF con-

tributes about 20%–25% to the magnitude of the large-scale atmospheric circulation (Bergman and Hendon 2000a).

However, several studies with atmospheric general circulation models (AGCMs) have shown that the simulated large-scale atmospheric circulation is very sensitive to atmospheric CRF. Without the atmospheric CRF of the tropical convective–cirrus clouds, the Hadley and Walker circulations are much weaker and the atmospheric thermodynamic structure is very different (e.g., Randall et al. 1989; Slingo and Slingo 1991; Sherwood et al. 1994; see more references in Tian 2002). Even a change of cloud parameterizations in AGCMs can significantly alter the Hadley and Walker circulations (Zender and Kiehl 1997). These AGCM studies indicate that atmospheric CRF may play a zeroth-order role in maintaining the tropical large-scale atmospheric circulation, which cannot be accounted for within the context of the Gill model (Bergman and Hendon 2000a; see more references and discussions in Tian 2002).

The primary objective of this study is to construct a simple dynamic model for the tropical atmospheric circulation following the moist thermodynamic formulation proposed by Neelin and Held (1987) and Gill (1982), that is, to refine the Gill (1980) model from a dry model into a moist one. The original Gill model views the tropical atmospheric circulation as the response of a dry atmosphere to dry diabatic processes such as latent heating and atmospheric CRF. However, the present model (a Gill–Neelin–Held model) treats the tropical atmospheric circulation as the response of a

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moist atmosphere to moist diabatic processes such as surface evaporation and atmospheric CRF. As first pointed out by Neelin and Held (1987), these moist diabatic processes relate to boundary conditions (surface heat fluxes) of the atmosphere. Thus, a simple moist model based on the moist static energy (MSE) equation should be the proper model for the tropical atmosphere circulation. Relying on this simple model and observed data of atmospheric CRF, we will elucidate the importance of atmospheric CRF in the large-scale tropical atmospheric circulation. However, this paper cannot provide a satisfactory causal explanation for the observed circulation because the strong coupling between the circulation, clouds, and surface evaporation is not incorporated in the present model. To fully understand the nature of the tropical circulation, a simple model with sea surface temperature (SST) as the only external forcing incorporating cloud–radiation–circulation feedback as well as wind–evaporation feedback is needed.

The plan for the rest of this paper is as follows. The model is formulated in section 2. Section 3 examines the role of atmospheric CRF in the circulation. Conclusions are summarized in section 4.

2. The model

The model atmosphere, which is a mixture of Gill (1980) and Neelin and Held (1987), describes perturbations around a basic state of rest with horizontally uniform density, temperature $[T_0(p)]$, geopotential $[\Phi_0(p)]$, and variable moisture limited to the lower troposphere $[q_0(x, y, p)]$. On a local pressure coordinate system, the steady-state-linearized governing equations for hydrostatic motions are written as follows:

The horizontal momentum equations are

$$\varepsilon_x u - fv = -\partial\Phi/\partial x, \qquad (1a)$$

$$\varepsilon_y v + fu = -\partial\Phi/\partial y, \qquad (1b)$$

where u and v represent zonal and meridional wind respectively, Φ is geopotential, and f is the Coriolis parameter. Here $\varepsilon_x u = g\partial\tau^x/\partial z$ and $\varepsilon_y v = g\partial\tau^y/\partial z$ are termed as Rayleigh friction (Gill 1980). The continuity equation for incompressible flow is

$$\partial u/\partial x + \partial v/\partial y + \partial\omega/\partial p = 0, \qquad (2)$$

where $\omega = dp/dt$ is the vertical pressure velocity. Following Gill (1982) and Neelin and Held (1987), the moist thermodynamic equation is

$$\omega\partial h_0/\partial p = -g\partial F/\partial p, \qquad (3)$$

where $F = F_R + F_L$ is the net heat flux due to radiation (F_R) and convection $[F_L = \overline{(L\omega'q')}]$. Here $h_0(x, y, p) = s_0(p) + Lq_0(x, y, p)$ is the MSE of the basic state and $s_0(p) = c_p T_0(p) + \Phi_0(p)$ is dry static energy. The sensible heating from surface sensible heat flux is neglected here because of its small magnitude and gradient.

Equations (1), (2), and (3) constitute our model Tropics. Considering the first baroclinic mode of the tropical troposphere (Gill 1980), the model reduces to a set of damped shallow-water equations:

$$\varepsilon_x u - fv = -\partial\Phi/\partial x, \qquad (4a)$$

$$\varepsilon_y v + fu = -\partial\Phi/\partial y, \qquad (4b)$$

$$\varepsilon_N \Phi + c_a^2(u_x + v_y) = -KQ, \qquad (4c)$$

where u, v, and Φ represent the low-level (e.g., 750 mb) horizontal wind and geopotential, respectively. Here $K = (R_d g)/(4c_p p_M)$ is constant, R_d is the gas constant for dry air, and c_p for the specific heat of dry air at constant pressure. Also, $p_M = (p_S + p_T)/2$ is the midtroposphere and $\Delta p = (p_S - p_T)/2$ is the half pressure depth of the troposphere; and p_S and p_T are pressures of the surface and the top of the atmosphere (TOA), respectively. By convention, the clear-sky radiative heating is parameterized using a uniform Newtonian cooling coefficient (ε_N) because of its small spatial gradient (Tian and Ramanathan 2002). Thus, the thermodynamic forcing Q in (4c) is defined as

$$Q = F_L(S) + C(A), \qquad (4d)$$

where $F_L(S)$ denotes surface evaporation, while $C(A)$ is atmospheric CRF.

In (4c), c_a denotes the phase speed of moist gravity waves and is defined as

$$c_a = (gh_e)^{1/2} = [(R_d\Delta h\Delta p)/(2c_p p_M)]^{1/2}. \qquad (4e)$$

Here Δh is referred to as "gross moist stability" (GMS; Neelin and Held 1987; Emanuel et al. 1994; Neelin and Zeng 2000; Zeng et al. 2000) and is defined as

$$\Delta h = h_{01} - h_{02} = \Delta s - Lq_{02}, \qquad (6f)$$

where $\Delta s = s_{01} - s_{02}$ is "gross dry stability" (Neelin and Held 1987). Here, s_{02}, q_{02}, and h_{02} are dry static energy, moisture, and MSE, respectively, of the lower troposphere, and s_{01} and h_{01} are those for the upper troposphere. Following Wang and Li (1993), we set $\Delta s = 27\,000$ J kg^{-1}, roughly corresponding to a dry gravity wave speed of 62 m s^{-1}. The low-level moisture (q_{02}) is calculated based on the formulas by Wang and Li (1993) and Neelin and Held (1987; see Tian 2002 for more details). The present model predicts that the wave speeds are about 10–25 m s^{-1}, which are consistent with the observed phase speeds of the equatorial gravity waves by Wheeler and Kiladis (1999) and are also consistent with the earlier theoretically predicted moist wave speeds by Emanuel et al. (1994), Neelin and Yu (1994), and Yu and Neelin (1994). However, we need to point out that the spatial variability of the GMS may be greatly overestimated by the simple vertical structure because the cloud-top height variations are not accounted here (see Yu et al. 1998). We also wish to point out here that the low-level moisture–convergence feedback (Webster 1981; Zebiak 1986) is represented by GMS in the present model (Neelin et al. 1998). Tian (2002)

TABLE 1. Basic parameters and their values used in the present model.

Symbol	Parameter	Standard value
p_T	TOA pressure	0 mb
p_S	Surface pressure	1000 mb
p_M	Midtroposphere pressure	500 mb
Δp	Half pressure depth of the troposphere	500 mb
ε_x	Zonal Rayleigh friction coefficient (4a)	1.5×10^{-5} s^{-1} (0.75 day)$^{-1}$
ε_y	Meridional Rayleigh friction coefficient (4b)	4.5×10^{-5} s^{-1} (0.25 day)$^{-1}$
ε_N	Newtonian cooling coefficient (4c)	1.0×10^{-6} s^{-1} (10 day)$^{-1}$
Δs	Gross dry stability (4f)	27 000 J kg^{-1}

found that GMS plays an important role in the large-scale tropical circulation and the results will be reported in a separated paper.

Solving the linear shallow-water equation (4) will give the linear response of lower-tropospheric wind and geopotential fields to the given thermodynamic forcing from atmospheric CRF and surface evaporation. The key parameters of this model are all taken from observations and summarized in Table 1.

3. Role of atmospheric CRF

Equation (4) has been used extensively to study the tropical atmospheric circulation, in particular by El Niño–Southern Oscillation (ENSO) community (see Philander 1990; McCreary and Anderson 1991; Neelin et al. 1998). The common way to deal with the forcing term given by (4d) is the following: surface evaporation is parameterized in terms of SST and atmospheric CRF is neglected, perhaps due to the difficulty in estimating atmospheric CRF. Thus, to our best knowledge, the relative role of atmospheric CRF and surface evaporation in this model has never been discussed in the existing literature. Sherwood et al. (1994) have examined a similar question in the context of a GCM, but not in a simple model like the present one. This study will fill this gap relying on recent estimates for atmospheric CRF by Tian and Ramanathan (2002). They infer atmospheric CRF using a combination of satellite radiation budget for CRF at the TOA and radiation models. The Comprehensive Ocean–Atmosphere Data Set (COADS) produced by da Silva et al. (1994) provides the observed
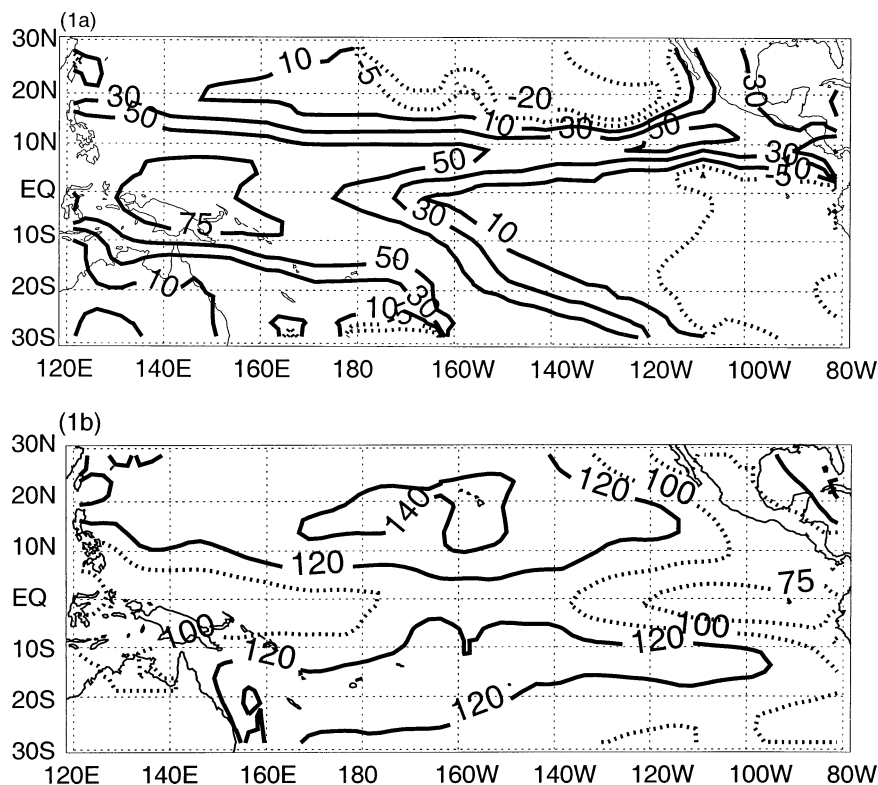

FIG. 1. The observed thermodynamic forcing of the present model (W m^{-2}): (a) atmospheric CRF and (b) surface evaporation.
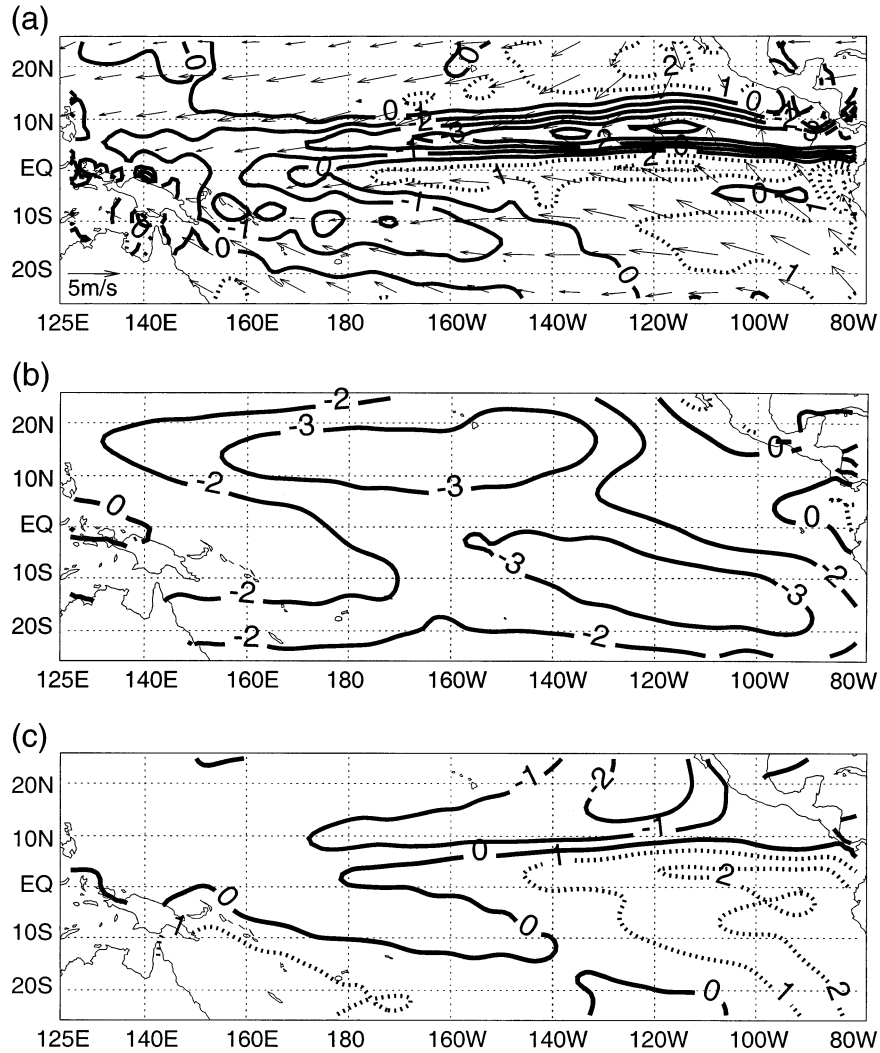
FIG. 2. The low-level horizontal winds (m s^{-1}) and divergence (10^{-6} s^{-1}) based on surface observations from COADS: (a) wind vector and divergence, (b) zonal wind, and (c) meridional wind.

surface evaporation. The atmospheric CRF and surface evaporation used in this study are shown in Fig. 1. The present model domain is 30°S–30°N, 100°–280°E with a spatial grid of 2.5° × 2.5° and time step of 150 s. The detailed numerical solution can be found in Tian (2002).

For comparisons, we first present annual-mean low-level horizontal winds and divergence in Fig. 2 based on the surface observations from COADS. The low-level winds from the surface observations are calculated employing a sinusoidal vertical structure following Davey and Gill (1987). As shown in Fig. 2, the low-level convergence zones are located in the intertropical convergence zone (ITCZ) and the South Pacific convergence zone (SPCZ), including the warm pool in the eastern equatorial Indian and the western equatorial Pacific Oceans. These low-level convergence zones correspond to the ascending branches of the Hadley and Walker circulations. The low-level divergence zones,

which form the descending branches of the Hadley and Walker circulations, are most predominant in the subtropics and the cold tongue. Accordingly, the dominant low-level flows in the tropical Pacific are the well-known trades blowing from the subtropics and cold tongue toward the ITCZ and the SPCZ. We wish to point out here that the Hadley circulation is fundamentally nonlinear (Held and Hou 1980), at least in its upper branch. Thus, the comparison of the model to the surface observation may not be adequate for the upper levels of the tropical circulation.

Low-level horizontal winds and divergence simulated by the present model when forced by just the observed annual-mean atmospheric CRF are shown in Fig. 3. Comparisons between Figs. 2 and 3 clearly show that when forced with the atmospheric CRF the model can reproduce the basic features of the Hadley and Walker circulation, such as the trades, the ascending branches
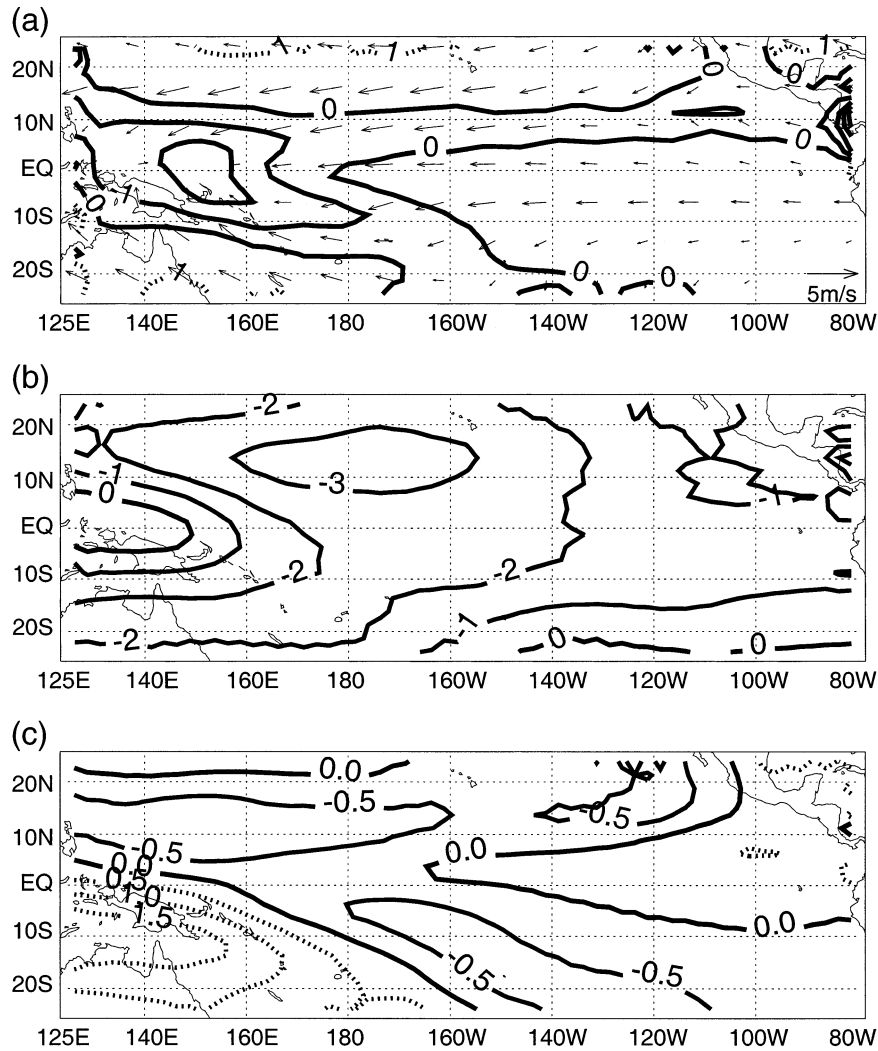
FIG. 3. The simulated low-level horizontal winds (m s^{-1}) and divergence (10^{-6} s^{-1}) when the present model is forced solely by the atmospheric CRF: (a) wind vector and divergence, (b) zonal wind, and (c) meridional wind.

in the ITCZ and the SPCZ, as well as the descending branches in the cold tongue and subtropics. The low-level convergence zone is also quantitatively captured by the model. For example, the magnitude of the low-level convergence is in the order of 10^{-6} s^{-1} in the ITCZ and SPCZ similar to observations. The present model can also accurately simulate the low-level zonal wind speeds. For example, the strong easterlies of around 3 m s^{-1} in the Pacific and the weak westerlies at the equator just north of Australia and around Central America are all quantitatively reproduced. The model can also qualitatively simulate the pattern of the meridional wind, such as the northerlies in the North Pacific and the southerlies in the South Pacific. However, we have to point out that the magnitude of model low-level convergence is overestimated in the warm pool and SPCZ, while it is underestimated in the ITCZ. The magnitude of the meridional wind is also underestimated. Further-

more, the northerlies in the eastern edge of SPCZ seem unrealistic. Considering the large errors in the estimated atmospheric CRF, these deficiencies are inevitable. These deficiencies may also be due to the lack of the boundary layer structure in the model (Lindzen and Nigam 1987; Wang and Li 1993).

Figure 4 presents the model-simulated low-level horizontal winds and divergence when the model is forced only by the observed annual-mean surface evaporation. The result indicates that, in the context of a moist model, the surface evaporation will force a circulation that tends to subside in the equatorial regions between 10°S and 10°N and rise in the subtropics. Accordingly, the Walker circulation totally disappears, and the Hadley circulation reverses. Thus, the observed spatial pattern of surface evaporation will drive a circulation that is far away from the observed circulation.

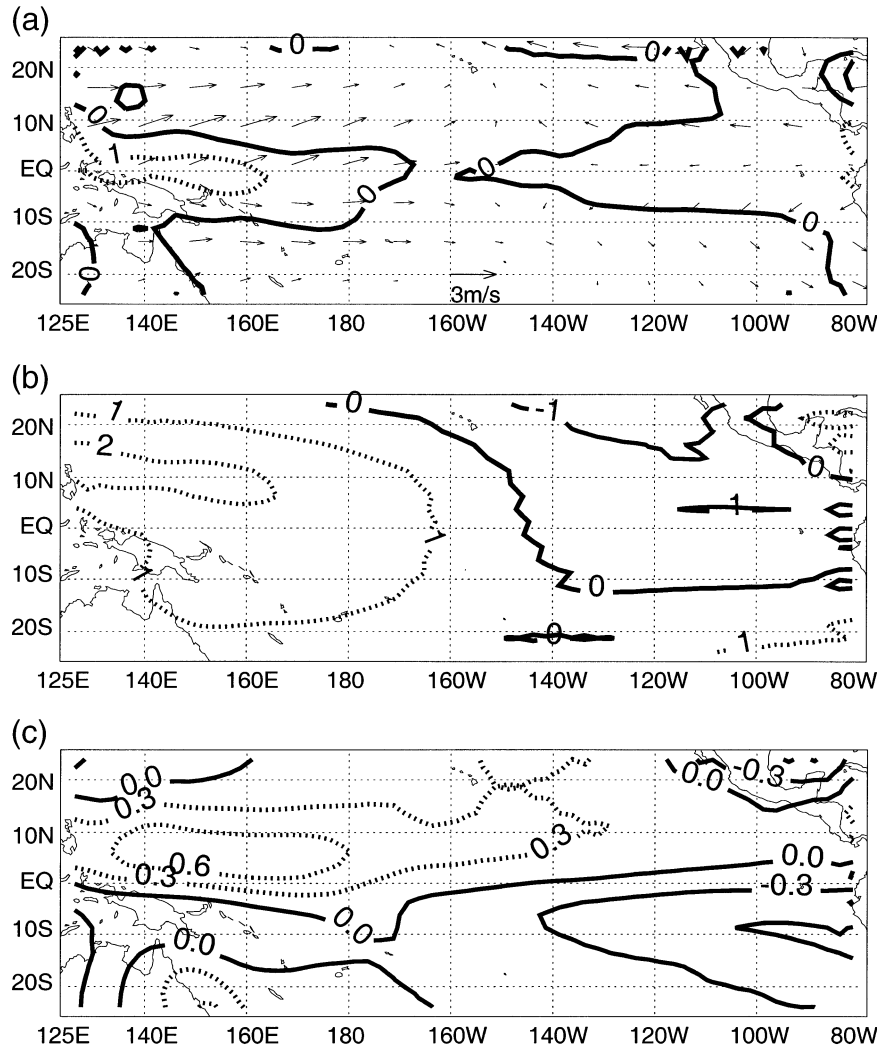A comparison between Figs. 3 and 4 clearly dem-

FIG. 4. The simulated low-level horizontal winds (m s^{-1}) and divergence (10^{-6} s^{-1}) when the present model is forced only by the surface evaporation: (a) wind vector and divergence, (b) zonal wind, and (c) meridional wind.

onstrates that the spatial variations of atmospheric CRF are more important than the spatial variations of surface evaporation for driving the circulation in the MSE budget. In fact, the evaporation does not even have the right spatial structure to drive the observed circulation. This result is consistent with several AGCM studies by, for example, Randall et al. (1989), Slingo and Slingo (1991), and Sherwood et al. (1994), which have shown that when the atmospheric CRF was removed from the AGCM, the simulated large-scale atmospheric circulation is very weak and is even eliminated completely. Thus, the present model may provide a better theoretical framework to understand the AGCM results than the original Gill model. However we point out the following important caveat. The diagnostic results presented here cannot be regarded as a causal explanation for the observed circulation because our model does not incorporate the strong coupling between the circulation, clouds, and surface evaporation. To fully understand the nature of the tropical circulation, a simple model with SST as the only external forcing incorporating cloud–radiation–circulation feedback as well as wind–evaporation feedback is needed.

4. Conclusions

In this study, a simple moist tropical atmospheric model is constructed based on the simple models of Gill (1980) and Neelin and Held (1987). The model describes the first baroclinic mode of the moist troposphere with variable GMS in response to the given thermodynamic forcing from atmospheric CRF and surface evaporation. This model has been used extensively to study the large-scale tropical atmospheric circulation, in particular by the ENSO community (see Philander 1990; McCreary and Anderson 1991; Neelin et al.

1998). However, to our best knowledge, no prior studies have discussed the relative role of atmospheric CRF and surface evaporation in the tropical circulation within the context of this model, perhaps due to the difficulty in estimating atmospheric CRF. The novelty of this study is to fill this gap relying on recent estimates for atmospheric CRF by Tian and Ramanathan (2002).

When the present model is forced solely by the observed atmospheric CRF, a quantitatively reasonable Hadley and Walker circulation is obtained, such as the trades, the ascending branches in the ITCZ and the SPCZ, as well as the descending branches in the cold tongue and subtropics. However, when the model is forced only by the observed surface evaporation, the Walker circulation totally disappears, and the Hadley circulation reverses. These results indicate that, in the context of a moist dynamic model, the spatial variations of atmospheric CRF are more important in terms of driving and maintaining the Hadley and Walker circulations than the spatial variation of surface evaporation.

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