Atmospheric Responses of Gill-Type and Lindzen–Nigam Models to Global Warming

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(Manuscript received 3 August 2010, in final form 11 May 2011)

ABSTRACT

The equatorial Pacific atmosphere responds differently to global warming in the Gill-type and Lindzen–Nigam models. Under an assumption of no change in the zonal sea surface temperature (SST) gradient in the Gill-type model, the Walker circulation is intensified in a warmer climate relative to current climatic conditions, while slightly weakened in the Lindzen–Nigam model. Furthermore, for more accurate derivation of the surface wind, the free atmosphere in the Gill-type model is combined with the atmospheric boundary layer. This modified Gill-type model actually produces weaker surface wind than the Gill-type model would, but the sensitivity of the Walker circulation to the warmer climate is similar to that obtained from the Gill-type model. These results may explain why the zonal gradient of equatorial Pacific SST during the twentieth century is observed to strengthen while the Walker circulation is not, even though they are dynamically linked.

1. Introduction

Unlike in the extratropics, the formulation of quasi-steady-state atmospheric circulation in the tropics is simple because of the relatively weak eddy activities and retains primary features, especially when it comes to longer time scale variability (greater than 1 week). The investigation of tropical climate phenomena may be facilitated by a simpler atmospheric model than that required to analyze the extratropics. The earlier Gill model (Gill 1980) and later Lindzen–Nigam model (Lindzen and Nigam 1987, hereafter referred to as LN) have been widely used for studies of tropical air–sea interactions, particularly the El Niño–Southern Oscillation (ENSO). A major difference between these two models is that the forcing term of the LN model belongs to the momentum equations, while the forcing term of the Gill model belongs to the thermodynamic equation. In the Gill model (Philander et al. 1984; Gill 1980; Hirst 1986; Zebiak 1986), the latent heating of the atmosphere (i.e., cumulus heating of the middle and upper troposphere) drives low-level winds proportional to SST anomalies (Hirst 1986) or related to surface heat flux anomalies (Zebiak 1982). Some models have further adopted the secondary feedback of the low-level moisture convergence (Zebiak 1986). On the other hand, the LN model calculates the boundary layer flow that is directly forced by surface temperature based on the strong relationship between horizontal temperature gradients and horizontal pressure gradients. In other words, local cumulus heating in the Gill model generates low-level flow, while the surface temperature gradient generates low-level flow in the LN model. However, Neelin (1989, hereafter N89) showed that the LN model could be transformed into the Gill model by neglecting the smaller terms [see Eq. (4) of N89], whereby atmospheric forcing, which is expressed as a form of SST gradient in the momentum equations of the LN model, is moved into the thermodynamic equation as a form of the SST itself (Zebiak 1982). The latent heating in the Gill-type model is usually parameterized to be proportional to SST, which is based on the temperature dependency of the evaporation rate.

Although the transformation performed by N89 is quite reasonable, there remains a difference regarding the actions of the inherent processes such that the low-level convergence drives cumulus convection in the middle–upper atmosphere (LN model) or vice versa (Gill-type model). The difference might be small over short time scales, but it cannot be disregarded in the context of a changing climate because changes in climate (especially mean SST) exert opposite effects on the two systems. For example, cumulus heating, that is, the forcing of the Gill-type model (Zebiak 1982, 1984) is primarily proportional to the amount of surface evaporation, which depends on the surface temperature. Since the evaporation rate...
increases with temperature, an SST anomaly that occurs over a warm sea surface provides more cumulus heating than does an equivalent anomaly that occurs over a relatively cool sea surface (e.g., Zebiak 1982). On the other hand, the pressure gradient or the low-level wind in the LN model results from the SST distribution. Because density depends on temperature, in the LN model the sea level pressure perturbation over a warmer surface is less sensitive to change in SST. Therefore, the atmospheric responses to changes in SST under warmer climate conditions are strengthened in the Gill model but suppressed in the LN model compared with those observed under cooler climate conditions. The purpose of this paper is to explore differences in responses to changing climate between the Gill-type model and the LN model, and to discuss the implications of these differences in the context of the debate over future changes of tropical Pacific SST.

2. Transforming the Lindzen–Nigam model into the Gill model

We include all of the terms that were neglected by N89 and therefore retain the original forms used in LN [see Eqs. (6c), (7c), and (10) of LN], which are

\[ eu' - fu' + g(2 - \frac{T_s}{T_0} + \alpha H_0/T_0) \frac{\partial h'}{\partial x} = \frac{g H_0}{2T_0} \left(1 - \frac{2\gamma}{3}\right) \frac{\partial T_s'}{\partial x}, \]  

(1a)

\[ ev' + fu' + g(2 - \frac{T_s}{T_0} + \alpha H_0/T_0) \frac{\partial h'}{\partial y} = \frac{g H_0}{2T_0} \left(1 - \frac{2\gamma}{3}\right) \frac{\partial T_s'}{\partial y} + \frac{gh'}{2T_0} \frac{\partial T_s}{\partial y}, \]  

(1b)

and

\[ \varepsilon T h' - H_0 \left(\frac{\partial u'}{\partial x} + \frac{\partial u'}{\partial y}\right) = 0, \]  

(1c)

where \( u' \) and \( v' \) are the horizontal velocities averaged through the depth of the boundary layer, \( h' \) is the perturbation height of the top of the boundary layer, and \( T_s \) is the surface temperature. The bar (prime) indicates the mean (deviation); \( H_0 \) (3000 m) is the mean depth of the boundary layer, \( T_0 \) (288 K) is the reference temperature for the linearization of the temperature dependence of density, and \( \varepsilon \) [(2 days)\(^{-1}\)] is the mechanical damping due to vertical diffusion of momentum and surface drag. Here \( \varepsilon T \) [(30 min\(^{-1}\)] is the inverse relaxation time of the adjustment of the boundary layer height, \( \alpha \) (0.003 K m\(^{-1}\)) is the lapse rate for temperature decay with height, and \( \gamma \) (0.30) is a parameter related to static stability (LN).

As seen in Eq. (1), the LN system is forced by pressure gradients proportional to the gradients of SST. By transformation

\[ h' = h' + \left(\frac{H_0}{2T_0}\right) \left[1 - \frac{2\gamma/3}{2 - \frac{T_s}{T_0} + \alpha H_0/T_0}\right] T_s', \]  

(2)

and

\[ \mathcal{H}_0 = (\varepsilon/\varepsilon_T) H_0, \]  

(3)

where the \( \mathcal{H}_0 \) is a boundary layer depth scaled by the ratio between the mechanical damping and the Newtonian cooling (or longwave cooling), and this system is converted to the Gill-type model in terms of the mathematical form:

\[ eu' - fu' + g[2 - \frac{T_s}{T_0} + \alpha H_0/T_0] \frac{\partial h'}{\partial x} = 0, \]  

(4a)

\[ ev' + fu' + g[2 - \frac{T_s}{T_0} + \alpha H_0/T_0] \frac{\partial h'}{\partial y} = \frac{gh'}{2T_0} \frac{\partial T_s}{\partial y}, \]  

(4b)

and

\[ \varepsilon T h' - \mathcal{H}_0 \left(\frac{\partial u'}{\partial x} + \frac{\partial u'}{\partial y}\right) = -\frac{\varepsilon H_0}{2T_0} \left[1 - \frac{2\gamma/3}{2 - \frac{T_s}{T_0} + \alpha H_0/T_0}\right] T_s'. \]  

(4c)

As seen in Eq. (4c), the forcing term appears in the height equation. The major differences between N89 and the model presented in this study are marked by the square brackets in the equations. The quantity in the square bracket is a function of the mean SST; thus, changes in mean SST cause the system to respond differently to the equivalent SST anomaly. If each of the square brackets stands for a unit, then the above system becomes identical to a system driven by N89 [see Eq. (4) in N89].

3. Two-box approximations of Gill-type and LN models

To assess the tropical air–sea coupling strength generated by both models, each system is converted into a two-box system. Like previous prototype equatorial atmosphere–ocean systems, each box is composed of the equatorial western and eastern basins. Temperature is defined for both boxes, and the surface wind is defined at the center of the basin (i.e., between the
boxes). The variable defined at the center is marked by the subscript \(c\).

To simplify the problem without losing the primary factor, the meridional components are neglected and the equatorial \(f\) plane, that is, \(f = 0\), is assumed. This is reasonable because in the tropics the magnitude of the meridional wind is smaller than that of zonal wind. Combining (4a) and (4c) yields

\[
\epsilon u' + \frac{gH_0}{\epsilon T} \lambda(T_s) \frac{\partial^2 u'}{\partial x^2} = \frac{gH_0}{2T_0} \left(1 - \frac{2}{3} \gamma\right) \frac{\partial T'}{\partial x},
\]

(5)

where \(\lambda(T_s) = 2 - \frac{T_s}{T_0} + \alpha H_0/T_0\) and \(T_s\) is assumed to be spatially constant. Equation (5) indicates the relationship between the zonal contrast of SST and low-level wind at the center referring to the trade wind. Since Gill (1980) did not specify the atmospheric heating, we started from the system of Eq. (A3) of Zebiak and Cane (1987). Under the same assumption used for the derivation of (5), the equation representing the relationship between the zonal contrast of SST and low-level wind in the Gill-type model becomes

\[
\epsilon u' - \frac{C_a}{\epsilon} \frac{\partial^2 u'}{\partial x^2} = \frac{K}{\epsilon} \alpha^* \exp \left(\frac{T_s - 30}{16.7}\right) \left(\frac{\partial T'}{\partial x} + \frac{T_s'}{16.7} \frac{\partial T}{\partial x}\right),
\]

(6)

where the secondary feedback of the low-level convergence, which was adopted in Zebiak and Cane (1987), is excluded, but the heating due to the local evaporation is remained. The constants are \(\epsilon [1(2\text{day})^{-1}]\), \(C_a (60 \text{ m s}^{-1})\), and \(\alpha^* (3.1 \times 10^{-2} \text{ m}^2 \text{ s}^{-3} \text{C}^{-1})\); the unit for the mean sea surface temperature inside the exponential function is Celsius. Converting parameter \(K\) is given by \(R_d \Delta p / (2C_p p_s)\), where \(p_s\) is the middle pressure level of the free atmosphere (=50 kPa) and \(\Delta p\) is the half-depth of the free atmosphere (=50 kPa). The atmospheric (cumulus) heating anomaly in Zebiak and Cane (1987) was derived from the linearized Clausius–Clapeyron equation representing the temperature-dependent change of local evaporation, that is, \(Q_E = \alpha^* T_s' \exp[(T_s' - 30)/16.7]\). More detailed derivations for the cumulus heating are found in Zebiak (1982, 1986).

Regarding the Gill-type model formula of Eq. (6), two issues had been clarified, and accordingly the model had to be modified, at least so as to perform a better simulation of the surface wind. The first issue is related to how effectively the cumulus convective heating in the middle troposphere drives the surface winds. Many studies (e.g., Battisti et al. 1999; Wu et al. 1999; Chiang et al. 2001) mentioned that the deep convective heating is not effective in driving surface winds; thus the usefulness of the Gill model, which captures the first baroclinic mode, may be limited for the free atmosphere (e.g., Battisti et al. 1999). The second issue is related to whether the convective cumulus heating is induced by the surface evaporation associated with change in SST. For example, Tompkins and Craig (1999) argued that the convection is insensitive to changing SST in the absence of large-scale flow. To resolve these issues but not miss the main propose of this study, following Wang (1988) and Wang and Li (1993), the free atmosphere is allowed to interact with the well-mixed atmospheric boundary layer. The steady-state atmospheric boundary circulations in this case are determined by the geopotential height at the top of the atmospheric boundary layer matching to the lower-level geopotential height in the free atmosphere. Furthermore, by including the convective heating by the surface moisture convergence, more realistic convective heating is calculated (e.g., Wang 1988; Kleeman 1991; Wang and Li 1993). The details in the model formula are shown in the appendix. After combining the atmospheric boundary layer model and the free atmosphere model, the equation for the relationship between the zonal wind in the atmospheric boundary layer and the surface temperature becomes

\[
\epsilon u_B - \frac{C_a}{\epsilon} \left[1 + \frac{\epsilon(p_s - p_e)^2}{2p_0 g K_D \Delta p}\right] \frac{\partial^2 u_B}{\partial x^2} = \frac{(p_s - p_e)}{p_0 g K_D \Delta p} \frac{\partial Q}{\partial x},
\]

(7)

where

\[
\frac{\partial Q}{\partial x} = -\frac{L_c}{\Delta p} \left[\frac{1}{2}(\overline{q}_3 - \overline{q}_1)(p_s - p_e) + (\overline{q}_e - \overline{q}_3)(p_s - p_e)(1 + a\delta T_s)\right] \frac{\partial u_B}{\partial x} + \frac{\partial Q_E}{\partial x},
\]

(8)

and \(p_s\) and \(p_e\) are 100 and 90 kPa, respectively. Since the zonal contrast of the change in the mean SST (i.e., \(\delta(T_s)/\delta x = 0\)) is not considered in this study, the second term of Eq. (8) can be neglected. Then, Eq. (7) becomes
\[ \varepsilon u_B = \frac{c^2}{\epsilon} \left[ 1 + \frac{\varepsilon(p_s - p_e)^2}{2 \rho g K_D \Delta p} - \frac{\varepsilon K^*}{C^2} M(\delta T_s) \right] \frac{\partial^2 u_B}{\partial x^2} \]
\[ = K^* a^* \exp \left( \frac{T_s - 30}{16.7} \right) \frac{\partial T'}{\partial x}, \tag{9} \]

where

\[ M(\delta T_s) = \frac{L_c}{\Delta p} \left\{ \frac{1}{2} \left( \overline{T}_3 - \overline{T}_1 \right) (p_s - p_e) \left[ 1 + \frac{2 \rho g K_D \Delta p}{\varepsilon(p_s - p_e)^2} \right] \right. \]
\[ + \left( \overline{\eta_e} - \overline{T}_3 \right)(p_s - p_e)(1 + a \delta T_s) \left\} \right. \]
\[ \right\} \]

and

\[ K^* = \frac{(p_s - p_e)}{\rho g K_D} K. \]

After some manipulation, the two-box approximations for Eqs. (5), (6), and (9), respectively, give

\[ u'_{c} = \frac{\left( g H_0 / 2 T_0 \right) (1 - 2 \gamma / 3)}{\varepsilon L / 2 - 8 g H_0 \lambda(T)/V_e L} \Delta T_s', \tag{10} \]
\[ u'_{c} = \frac{K_\alpha^* \exp(\left( T_s - 30 \right)/16.7)}{(\varepsilon L / 2 + 4 C_\alpha^2 / L)} \Delta T_s', \tag{11} \]

and

\[ u'_{c} = \left\{ \frac{K_\alpha^* \exp(\left( T_s - 30 \right)/16.7)}{(\varepsilon L / 2 + 4 C_\alpha^2 / L)} \left[ 1 + \frac{\varepsilon(p_s - p_e)^2}{2 \rho g K_D \Delta p} - \frac{\varepsilon K^*}{C^2} M(\delta T_s) \right] \right\} \Delta T_s', \tag{12} \]

where \( L \) (15 000 km) indicates the ocean basin length, \( \Delta T_s' \) is the SST difference between the eastern basin and western basin (i.e., SST in the eastern box minus SST in the western box), and \( u'_{c} \) is the surface zonal wind defined in the middle of two boxes. The zonal gradient of the mean SST in Eqs. (6) and (9) is neglected because of \([T_s'_{\text{cont}}]/16.7 \sim 0(0.01)\). For \( T_s = 300 \) K the ratios between the zonal wind and the zonal SST gradient \( (u'_{c}/\Delta T_s') \) are 2.64 (LN model), 3.06 (Gill-type model), and 1.60 (modified Gill-type model or Wang model). Thus, the strength of the zonal wind with respect to zonal SST gradient is largest in Gill-type model, in the middle in the LN model, and smallest in the modified Gill-type model. For example, when the zonal SST gradient between the western and eastern Pacific is \(-3 \) K, a typical value in the equatorial Pacific, the surface zonal wind is \(-7.92 \) m s\(^{-1}\) in the LN model, \(-9.17 \) m s\(^{-1}\) in the Gill-type model, and \(-4.18 \) m s\(^{-1}\) in the modified Gill-type model, which are overall consistent with the observed equatorial surface zonal wind.

Equations (10)–(12) again indicate an air–sea coupling process from the atmospheric point of view, meaning that the increase of the zonal SST gradient intensifies the low-level zonal wind over the equatorial central Pacific and the decrease of the zonal SST gradient reduces the same wind. The coupling strength is proportional to the value of combined parameters on the right-hand side of each equation. These parameters are functions of the mean SST, and thus the changing climate leads to changes in the air–sea coupling strength. To see how much the mean SST modifies the air–sea coupling, here we compare the low-level wind at a climate state \( (T_1) \) with that at another climate state \( (T_2) \) in the

LN model:

\[ \begin{align*}
    u'_{c}(T_2) &= \frac{1 - 16 g H_0 \lambda(T)/V_e L^2}{1 - 16 g H_0 \lambda(T_2)/V_e L^2}, \\
    u'_{c}(T_1) &= \frac{1 - 16 g H_0 \lambda(T)/V_e L^2}{1 - 16 g H_0 \lambda(T_1)/V_e L^2},
\end{align*} \tag{13} \]

Gill-type model:

\[ \begin{align*}
    u'_{c}(T_2) &= \frac{\exp[\left( T_2 - 30 \right)/16.7]}{\exp[\left( T_1 - 30 \right)/16.7]}, \\
    u'_{c}(T_1) &= \frac{\exp[\left( T_2 - 30 \right)/16.7]}{\exp[\left( T_1 - 30 \right)/16.7]},
\end{align*} \tag{14} \]

and the modified Gill-type model with a boundary layer (Wang model):

\[ \begin{align*}
    u'_{c}(T_2) &= \frac{\exp[\left( T_2 - 30 \right)/16.7]}{\exp[\left( T_1 - 30 \right)/16.7]} \left\{ \frac{\varepsilon L / 2 + 4 C_\alpha^2 / L}{\varepsilon L / 2 + 4 C_\alpha^2 / L} \left[ 1 + \frac{\varepsilon(p_s - p_e)^2}{2 \rho g K_D \Delta p} - \frac{\varepsilon K^*}{C^2} M(T_1 - T_r) \right] \right\}, \\
    u'_{c}(T_1) &= \frac{\exp[\left( T_2 - 30 \right)/16.7]}{\exp[\left( T_1 - 30 \right)/16.7]} \left\{ \frac{\varepsilon L / 2 + 4 C_\alpha^2 / L}{\varepsilon L / 2 + 4 C_\alpha^2 / L} \left[ 1 + \frac{\varepsilon(p_s - p_e)^2}{2 \rho g K_D \Delta p} - \frac{\varepsilon K^*}{C^2} M(T_2 - T_r) \right] \right\},
\end{align*} \tag{15} \]

for which \( T_r \) is the reference mean SST.

The ratios (\%) of zonal winds with respect to the different mean SSTs obtained by using Eqs. (13)–(15) are shown in Fig. 1, which exactly indicates \( (u'_{c}(T_2)/u'_{c}(T_1) - 1) \times 100(\%) \). For the calculation, the initial climate state of \( T_1 = T_r = 300 \) K is used. As seen in the
Figure 1. Change (%) of the surface (low level) zonal wind in the central Pacific with respect to the mean SST change (from $T_1$ to $T_2$) obtained from the two-box approximation of each model type: the Lindzen–Nigam model (LN), Gill-type model (Gill), and the modified Gill-type model (Wang).

The zonal wind is largely intensified in both Gill-type and modified Gill-type models, while slightly weakened in the LN model. For example, when the climate state warms to $T_2 = 305$ K, the surface (low level) winds with respect to the equivalent SST gradient are reduced by 3% in the LN model and increased by 34% in the Gill-type model (36% in the modified Gill-type model). In other words, as the climate warms, the sensitivity of the low-level wind to changes in the SST gradient (which indicate the large-scale air–sea coupling strength) increases in the Gill-type model but slightly decreases in the LN model. For the modified Gill-type model (Wang model), the air–sea coupling strength is modified not only, like in the Gill-type model, by change in the surface evaporation rate due to change in the mean SST but also change in the atmosphere moisture content associated with the mean temperature. However, the sensitivity of the low-level wind to changes in the SST gradient of the modified Gill-type model is similar to that of the Gill-type model. This is because the mixing ratio in the boundary layer is little changed as the mean SST increases. On the whole, it is expected that these systems (i.e., LN, Gill-type and modified Gill-type models) should have different responses to global warming, referring to the driving mechanism of the low-level convergence in the tropical atmospheric boundary layer.

The scaled boundary layer depth $H_0$ in Eq. (4) [or equally can be used as equivalent depth in the form of the Gill-type model like Eq. (4)] is about 32 m with $H_0 = 3000$ m, which corresponds to 17.7 m s$^{-1}$ of the gravity wave speed (or equatorial Kelvin wave speed). This speed is close to the observed, which is about 12–50 m s$^{-1}$ depending on the convectively coupled or non-convectively coupled (Banthier and Wallace 1996; Milliff and Madden 1996; Wheeler and Kiladis 1999), but the boundary layer depth assumed in the LN model is still deep. Therefore, we further check the sensitivity of the models with respect to the boundary layer depth. Moreover, the CGCMs under the global warming scenario produced a more stably stratified atmosphere (i.e., increase of dry static stability), which possibly causes a slowdown of the large-scale atmosphere circulation (Knutson and Manabe 1995; Chou and Chen 2010). The stabilized stratification effect actually can be adopted into a simple atmosphere model by changing the equivalent depth (e.g., Zheng et al. 2010). This is because the equivalent depth is directly related to the static stability. For example, the free adiabatic Kelvin wave phase speed is $c = (gh_c^{1/2})^{1/2}$, where $h_c$ is the equivalent depth, $m_z = 2\pi L_z$, the vertical wavenumber, and $N_h$ the Brunt–Väisälä frequency (Fuchs and Marki 2007); thus the equivalent depth is linearly proportional to the static stability for $m_z = \text{const}$.

To check the sensitivity of models to the equivalent depth, the ratios, like in Fig. 1, are computed using the same equations as used here but with a different boundary layer depth. For the LN model, the range of $H_0$ is taken between 50 and 3000 m, and for both Gill-type and modified Gill-type models, $C_a$ is chosen as 9.6 – 60 m s$^{-1}$. The term of $C_a$ in the Gill-type model was retained in Eq. (11) so that the zonal wind amplitude is changed as the gravity wave speed changes. However, $C_a$ disappears in the ratio, Eq. (14), and thus the ratio cannot be changed. The ratios for the modified Gill-type model were also little changed for the different $C_a$. Therefore, the ratios of zonal wind response to the different mean SSTs in both Gill-type and modified Gill-type models are merely influenced by change in the gravity wave speed (not shown here). On the other hand, some modifications in the LN model with respect to the boundary layer depth are observed as shown in Fig. 2. As seen in Fig. 2, for the shallow boundary layer depth the ratio (%) of zonal wind speed becomes smaller, while the ratio increases for a deeper boundary layer depth. Thus, the deep boundary layer more effectively reflects the mean temperature change in its zonal wind response.

4. Discussion and summary

Whether the tropical Pacific response to future global warming will resemble El Niño or La Niña is an ongoing debate (Cane et al. 1997; Vecchi and Soden 2007; Latif and Keenlyside 2009; Karnauskas et al. 2009; DiNezio et al. 2009; Collins et al. 2010). Observational reconstructions
of tropical Pacific SST during the twentieth century showed a long-term trend resembling La Niña (Cane et al. 1997; Hansen et al. 2006; Karnauskas et al. 2009; Compo and Sardeshmukh 2010). This observed trend is likely due to the large compensation of radiative heating by the upwelling in the eastern Pacific known as the ocean dynamical thermostat mechanism (Clement et al. 1996; Cane et al. 1997; DiNezio et al. 2009). At the same time, weakened zonal atmospheric overturning circulation (i.e., Walker circulation) due to global warming has been suggested based on the theory of mass and energy balances of the atmosphere (Held and Soden 2006; Vecchi and Soden 2007). Vecchi et al. (2006) provided more evidence to support the theory of a weakened zonal atmospheric overturning circulation by analyzing annual-mean sea level pressure in the Indo-Pacific sector. Recently, Karnauskas et al. (2009) showed that the observed zonal SST gradient along the equatorial Pacific Ocean has been strengthened, while the sea level pressure gradient has not. Zonal contrasts of the SST and Walker circulation over the equatorial region are dynamically linked; thus, observed atmospheric and oceanic trends in the tropical Pacific are seemingly inconsistent with each other. However, Xie et al. (2010) argued that the global warming due to the increasing greenhouse gas concentration and El Niño differ in mechanism such that the zonal SST pattern under the global warming along the equator is formed through a dedicated balance among upwelling damping, thermocline feedback, zonal advection, and evaporative cooling. As suggested by Seager and Murtugudde (1997), from an oceanic point of view, increases in the ocean’s thermal stratification can maintain the strong zonal SST gradient without strengthening the trade wind. Interestingly, a CGCM under the global warming scenario showed that the ocean dynamical thermal transport in the eastern equatorial Pacific becomes negligible because of a cancellation between the upwelling damping and thermocline feedback effect (Xie et al. 2010). On the other hand, as shown from the atmospheric point of view, the air–sea coupling can be either reduced or increased, depending on how the atmosphere responds to changes in SST as the mean SST increases. If the atmospheric response follows the LN system, the zonal atmospheric overturning circulation is not necessary to accompany strengthening of the zonal SST gradient as tropical warming progresses (note that the change in static stability modifies the atmosphere response as well, as shown in the previous section). However, if the atmospheric response follows the Gill-type system (including the modified Gill-type), the zonal atmospheric overturning circulation can be further intensified under warm climate conditions—more than it would be under cold climate conditions. Therefore, the atmospheric dynamical process (especially in the boundary layer) that rules out the atmospheric response to changes in SST must be a key “knob” to determine tropical Pacific climate states that are relevant to future global warming.

Since we use a simple atmospheric model, in the following the limitations of this study are discussed. First, the approach taken here was a forced response, namely the atmospheric response to a fixed SST forcing. However, it is well known that the tropical atmosphere and ocean are tightly linked to each other, for example, the Bjerkness feedback (Bjerknes 1969). Thus, the atmospheric response to global warming must be determined though air–sea interaction including the dynamical (i.e., the upwelling damping, thermocline feedback, and zonal advective feedback) and thermodynamical feedbacks (i.e., evaporative cooling, longwave radiative cooling) existing in the tropical as well as the extratropical region (DiNezio et al. 2009; Seager et al. 2010; Xie et al. 2010). Also the cloud–albedo feedback must be taken into account. Second, the final products of this study such as Fig. 1 were computed under the assumption of a non-meridional velocity and \( f \) plane. Under these approximations, the shallow-water model may have gravity wave components (including equatorial Kelvin waves) but not Rossby wave components, which may be somewhat problematic in the free wave solutions. However, the forced response, as in this study, may not result in significant error. The simple tropical atmosphere–ocean coupled models with and without a rotational component in the equatorial beta plane produced a qualitatively similar near-equatorial wind response to a given SST (e.g., An 2000; An and Kang 2001). This is because the
near-equatorial response is dominated by the Kelvin wave component rather than Rossby wave component owing to the weak Coriolis effect. Therefore, the exact quantity of the wind response may not be the same, but a qualitative comparison, as in this study, is still valid. Finally, again because we dealt with the atmosphere in a simple context and focused on one aspect, some parameter values and formulation may be arguable. Nevertheless, this study addressed how the surface evaporation rate change, the atmospheric mixing ratio change, and the atmosphere density dependency change with respect to change in the mean SST will modify the low-level zonal wind response to zonal SST contrast. The relative importance of each effect and the understanding of role of each effect are the main values of this study.

Acknowledgments. The author thanks J. Choi and S.-H. Im for drawing figures. This work was supported by the National Research Foundation of Korea Grant funded by the Korean Government (MEST) (NRF-2009-C1AAA001-2009-0093042).

APPENDIX

Gill-Type Model with the Atmospheric Boundary Layer

The boundary layer model is developed following Wang (1988). Started from equations for a steady, barotropic, neutral boundary flow, the vertically averaged horizontal winds in the boundary layer ($u_B$ and $v_B$), defined between the surface pressure $p_s$ and the pressure at the boundary layer top $p_e$, and the vertical $p$ velocity at the top of the boundary layer, $\omega_e$, are given by

$$-fu_B = -\frac{\partial \phi}{\partial x} - \frac{\rho_g g}{p_s - p_e} K_D u_B,$$

(A1)

$$fu_B = -\frac{\partial \phi}{\partial y} - \frac{\rho_e g}{p_s - p_e} K_D v_B,$$

(A2)

and

$$\omega_e = (p_s - p_e) \left( \frac{\partial u_B}{\partial x} + \frac{\partial v_B}{\partial y} \right),$$

(A3)

in which

$$u_B = \frac{1}{p_s - p_e} \int_{p_s}^{p_e} u dp,$$

(A4)

and $K_D (=0.0301)$ is the surface drag, $\rho_e$ (1.0 kg m$^{-3}$) is the air density in the boundary layer, and $\phi$ is the geopotential of the baroclinic component of the free atmosphere. Note that the barotropic components of the free atmosphere are neglected because the convective heating cannot induce barotropic components (e.g., Gill 1980). The $u_B$ and $v_B$ can be assumed as the surface winds (Wang 1988). In this system, the winds and geopotentials in the free atmosphere are driven by convective heating, and change in the geopotential leads to perturbation winds in the boundary layer. Near the equator, the pressure gradient force is likely balanced with the surface drag forcing. Therefore, Eq. (A1) becomes

$$u_B \approx -[(p_s - p_e)(\rho_e g K_D)] \frac{\partial \phi}{\partial x}.$$

The zonal momentum, continuity, and thermodynamic equations for the steady-state lower-level free atmosphere, fetched from Eqs. (3.4)–(3.9) of Wang (1988) and after some manipulations under the assumption of near-equator and very weak meridional winds, become

$$\varepsilon u \approx \frac{\partial \phi}{\partial x},$$

(A6)

$$\frac{\partial u}{\partial x} + \frac{1}{2\Delta p}(\omega_e - 2\omega_2) = 0,$$

(A7)

$$\varepsilon \phi + C_a^2 \frac{\partial u}{\partial x} \approx -\frac{C_a^2}{2\Delta p} \omega_e - KQ,$$

(A8)

and

$$Q = \frac{L_c}{\Delta p} [\omega_2 (\bar{q}_3 - \bar{q}_1) + \omega_e (\bar{q}_e - \bar{q}_3)] + Q_E,$$

(A9)

where subscripts 1, 2, and 3, indicate the levels in the free atmosphere that have equal pressure depth; $C_a$ is the gravity wave speed of the gravest baroclinic mode; the first two terms of $Q$ indicate the convective heating associated with the precipitation rate due to moisture convergence into the atmospheric column in the free atmosphere and atmospheric boundary layer, respectively; and the third term does that due to the surface evaporation used in Zebiak and Cane (1987) or also in Eq. (6). Here the middle pressure level of the free atmosphere, $p_2$, is 50 kPa, and the half-depth of the free atmosphere, $\Delta p$, is 40 kPa. Combining Eq. (A1) and (A6), we have $u_B = [\varepsilon (p_s - p_e)/(\rho_e g K_D)]u$. For a pressure depth of the boundary layer of 10 kPa, we have $u_B \approx 0.2u$, indicating that the zonal wind in the boundary
layer is about 20% of the zonal wind in the lower troposphere (e.g., Wu et al. 1999).

To consider the mean humidity change with respect to change in the mean temperature, the mean vapor mixing ratios in both the atmospheric boundary layer and lower layer of the free atmosphere are equally parameterized so as to depend on the mean surface temperature change ($\delta T_s$)—such that the difference of the mean mixing ratio between the boundary layer and the lower layer of the free atmosphere becomes $(1 + a \delta T_s)(\overline{q}_e - \overline{q}_3)$, where $a = 0.052$ (Xie et al. 2010). This is because the rational change in the humidity is highly correlated with the local SST (Xie et al. 2010). It has been assumed that the change in specific humidity is proportional to change in the mixing ratio, and the change in $(\overline{q}_3 - \overline{q}_1)$ associated with the mean SST is neglected because of its relatively small value.

REFERENCES


