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"A Consistent Model for the Large-Scale Steady Atmospheric Surface Circulation in the Tropics "

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A Consistent Model for Large the Steady Atmospheric Surface Circulation in the Tropics

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ABSTRACT

We present a new model of the tropical surface circulation, forced by changes in sensible heat and evaporative flux anomalies that are associated with prescribed sea surface temperature (SST) anomalies. The model is similar to the Lindzen and Nigam (L/N) boundary layer model, also driven by the above flux anomalies, only here the boundary layer is assumed well mixed and capped by an inversion. Hence, the model reduces to a two-layer, reduced gravity system. Furthermore, the rate of exchange of mass across the boundary layer/free atmosphere interface is dependent on the moisture budget in the boundary layer. When convection is diagnosed to occur, detrainment operates on the time scale associated with the life-cycle of deep convection, approximately eight hours. Otherwise, the detrainment is assumed to be associated with the mixing out of the stable tropical boundary layer which has a time scale of about one day. The model provides a diagnostic estimate of the anomalies in precipitation. However, we have assumed that the latent heat is released above the boundary layer and it drives a circulation that does not impact the boundary layer.

We discuss the inconsistencies between the formulations of both the Gill/Zebiak (G/Z) and Lindzen/Nigam models, and the values of several of the parameters that are required for these models to achieve realistic solutions (circulations). Then, the new reduced gravity boundary model equations are re-written in the form of the G/Z and L/N models. Using realistic values for the parameters in the new model geometry, we show that the constants combine in the re-written equations to produce the unphysical constants in the G/Z and L/N models; hence, the reason for the apparent success of these models.

1. Introduction

Over the last two decades, there were a considerable number of studies published in which investigators presented models of the steady tropical atmosphere response to sea surface temperature (SST) anomalies. These studies, summarized below, were primarily adapted from the classical work of Adrian Gill on the steady response of the tropical atmosphere to heating.

In his original paper, Gill (1980) proposed that the major features of the tropical atmospheric circulation could be explained by the linearized equations of motion, damped by mechanical friction and radiational cooling, and forced by deep tropical convection (Fig. 1a). Furthermore, Gill assumed that the vertical distribution of the convective forcing was such that the atmosphere response was given by an assumed gravest baroclinic mode, confined to the troposphere. The equations for the Gill model reduce to that of a two-layer atmosphere driven by latent heat release Q_o in convective towers which is assumed to be distributed equally in each layer. The equations, written here for the surface layer, are:

$$\epsilon_D \, \mathbf{u} \; - \; \beta \, \mathbf{y} \, \mathbf{k} \, \mathbf{x} \, \mathbf{u} \; = \; - \nabla \, \Phi \tag{1}$$

$$\epsilon_D \, \Phi \; + \; {c_a}^2 \nabla \cdot \mathbf{u} \; = \; -\mathbf{Q_o}, \tag{2}$$

where $\mathbf{u}$ is the mass weighted horizontal velocity averaged through the lower layer, Φ is the surface geopotential, c_a the atmospheric wave speed ($c_a = ND/\pi$, where N is the Brunt-Vaisalla frequency and D the troposphere thickness), βy is the local Coriolis parameter and ${\epsilon_D}^{-1}$ is the thermal and mechanical dissipation time. The variables $\mathbf{u}$, Φ and Q_o are taken to be perturbations about a prescribed state. **Figure 1.**

Investigators subsequent to Gill used this model, or modifications of this model, to examine the effects of SST anomalies in the Pacific Ocean during El Niño/Southern

Oscillation (hereafter, ENSO) on the surface winds (Gill and Rasmusson (1984), Zebiak (1986), Tian (1988)). Gill and Rasmusson (1984) used the *pattern* of the outgoing longwave radiative anomaly from the 1982-83 ENSO event as a proxy for heating Q_o and compared the model response to the observed circulation anomalies at 850mb in the tropical Pacific. They noted that the pattern of the response was consistent with the observed circulation pattern.

Gill and Rasmusson also noted that the convective (heating) anomalies were intimately tied to SST anomalies (see also, Webster, 1981). However, the prescription of Q_o from SSTs is not obvious, since in the 'Gill model' the convective forcing modifies the surface winds, whose convergence determines the heating. Zebiak (1982,86) explicitly included this effect by breaking up the heating Q_o into an evaporative component, Q_{evap}, and a CISK-like component resulting from a change in the surface mass convergence, Q_{conv}, due to the anomalous surface circulation:

$$Q_o = Q_{evap}(T', \overline{T}, rh) + Q_{conv}(\overline{q}, U', \overline{U}), \tag{3}$$

where q is the mixing ratio, rh is the relative humidity (assumed constant), and overbars and primes denote climatological means and anomalies, respectively. Equations (1) through (3) are then solved by iteration. Zebiak (1986) presented the surface circulation which resulted when the Rasmusson and Carpenter (1982) (hereafter RC) composite ENSO SST anomaly was used to calculate Q_{evap} and the Gill model was forced by Q_o as defined in Eq (3). He compared the model produced surface winds to the RC composite ENSO anomalies and noted the convergence feedback term improved the amplitude and pattern of the model response. Tian (1988) modified Q_{conv} to incorporate the effects of non-uniform fields of surface *moisture* convergence, $Q'_{conv}((q\mathbf{u})', \overline{q\mathbf{u}})$ rather than mass convergence. Tian (1988) found the model circulation anomalies to be slightly improved over that forced by mass convergence. Davey and Gill (1987) used the depth integrated moisture equation to determine Q_o from an estimated precipitation rate.

Despite the apparent success of the Gill model in reproducting the surface circulation anomalies during ENSO, some fundamental problems are known to exist with this model. As pointed out by Sardeshmukh and Hoskins (1985), the upper level vorticity balance in the tropics is inherently nonlinear. The Gill model is applied to the upper level flow and is linear; it can not reproduce this vorticity balance. Another problem with the Gill model is the need for an extremely short damping time (1-2 days) to horizontally confine the response (realistic damping affords a global response to isolated heating) and produce realistic meridional winds. It is also well known that the heating due to penetrative convection does not extend to the surface. Thus, the surface circulation should be extremely sensitive to the cloud base level (Battisti, 1984; unpublished manuscript; see also Wang (1988)) and the lower level in the Gill model will suffer extreme shear[1]. The decoupling of the observed surface and 850 mb flow field anomalies during ENSO is documented by Deser (1989).

We have discovered another problem with the Gill model which is ultimately related to the intricate coupling between the surface circulation and the upper level circulation. A re-examination of the scaling of Zebiak's evaporation term Q_o in Eq (3) has yielded an error, leading to surface latent heating anomalies from evaporation in excess of 75 Wm^{-2} for a 2°C temperature anomaly. These flux anomalies, averaged over a typical ENSO event, are four to five times larger than those observed for an SST anomaly of this size (see, e.g., Deser, 1989). When the correct scaling for Q_{evap} is incorporated into Eq (3), the resulting surface circulation anomalies forced by SST anomalies are

[1]The circulation in the tropics is constrained through a required balance between vertical motion and diabatic heating only when the heating is strong, i.e. in convective regions (Holton, 1992). Pressure gradient forces can significantly modify this balance in weakly diabatic regions.

insignificant[2].

Together, these problems force us to conclude the "Gill model", when conceived of as a model for tropospheric scale circulations, can not be used to generically simulate the tropical tropospheric circulation anomalies, especially when forced indirectly via surface heat fluxes.

Lindzen and Nigam (1987) (hereafter LN) have taken an alternative approach to modeling the tropical surface circulation anomalies. They assume SST anomalies produce sensible heat and evaporative fluxes which, in turn, produce virtual temperature anomalies, $T_v'(x, y)$ that are rapidly mixed into a confined boundary layer, producing surface pressure gradients that drive a surface circulation (Fig. 1b). They further assume that any surface convergence is everywhere rapidly vented by convection: LN set thermal damping time, ${\epsilon_T}^{-1}$, based on the approximate time scale for the *development* of convection - 30 minutes. An implicit assumption in the LN model, and contrary to the "Gill model", is that the upper-level convection plays a minor role in forcing the low-level wind field. Instead, the boundary layer anchors the location of the convection to the underlying SST anomalies and the upper-level only acts as a reservoir for the lower atmosphere heat and momentum perturbations.

While we basically agree with the LN model, there are some fundamental problems with the model. First, on time scales longer than ${\epsilon_T}^{-1}$, the tropical atmosphere is not everywhere convecting. Where the tropical atmosphere is not convecting, the boundary layer should relax on a time scale associated with mechanical damping or entrainment, which occurs on time scales on the order of ${\epsilon_D}^{-1} \simeq 1 - 2$days. Second,

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[2]Kleeman (1991) has developed a parameterization for Q_o in the Gill model that is virtually identical to that in Tian (1988). He also found that evaporative anomalies did not drive a significant circulation and was forced to prescribe a large, *ad hoc* forcing to achieve reasonable amplitude circulation anomalies.

Neelin (1989) has shown that the boundary layer convergence in the LN model, which is indicative of convective anomalies in the LN formulation, is proportional to the SST anomaly. However, the observed SST anomalies are not well correlated with anomalies in convection (see, e.g., Deser, 1989). Finally, the convective adjustment time ϵ_T^{-1} required for the LN model to achieve physically realistic solutions is O(30 minutes), which is unjustifiable. The large-scale relaxation of the tropical boundary layer will either be on time scales of the *life cycle* of the convection (approximately eight hours) or, in regions of strong subsidence, on the time scale associated with entrainment processes ($\simeq$ 1 day).

In this paper, we will incorporate the effects of the density discontinuity at the top of the boundary layer on the thermally induced pressure anomaly within the boundary layer. In this way, physically realistic and consistent solutions will be obtained and the inconsistencies noted above in previous models will be reconciled. We will show that by differentiating between convective and non-convective regions, a boundary layer model will better reproduce both the tropical surface circulation during ENSO and indicate the regions of anomalous convective activity.

The paper proceeds as follows. In section 2, we briefly review the model of LN, and develop a modified reduced-gravity model based on a moisture budget criterion. Also in this section is a brief discussion of the relationship between the depth integrated transport and the surface winds in this model. A critique of the new model in relation to previous studies is presented in section 3. A summary and discussion is presented in section 4.

2. The Model Development

2.a. The Boundary Layer Model

Lindzen and Nigam, in developing their boundary layer model, noted the following. Over the oceans, the lower tropical troposphere is usually well mixed due to buoyant convection originating from the surface. They argued that the mixed layer, which includes the surface layer and in places the stratocumulus/cumulus cloud layer, extends to typically two or three kilometers, where it is capped by a strong temperature inversion brought about by subsiding, drier air (see, e.g., Riehl (1979), Sarachik (1985)), particularly in the regions of the undisturbed trades. Even on spatial scales that are large compared to those of deep convection, it is likely that the tropics will also be characterized by a boundary layer capped by a substantial inversion.

Lindzen and Nigam, using the FGGE data, further noted that the *eddy* temperature field, $T_v'(x,y)$, defined as the deviation from the zonal mean, is vertically well-correlated in the lower troposphere below 700 mb. Hence, they assumed the following simple expression for the eddy temperature field in the lower troposphere:

$$T(x,y,z) = \overline{T}(y,z) + T_v'(x,y)\left(1 - \frac{\gamma}{H_b}z\right), \tag{4}$$

with

$$\overline{T}(y,z) = \overline{T}_s - \alpha z, \tag{5}$$

where $\overline{T}_s$ is the undisturbed zonal mean surface temperature, α is the undisturbed lapse rate, H_b is a reference height (taken by LN to be 3 km, or the approximate height of the 700 mb surface), and γ is an $O(1)$ constant controlling the diminution with height of the expression of the surface eddy temperature field. Here, as in LN, all temperatures refer to the *virtual* temperature.

In the LN model the eddy temperature field gives rise to a pressure gradient that defines the surface layer circulation. The horizontal momentum equation for the mass-weighted average flow in the boundary layer is

$$\epsilon_D \mathbf{U}_b \;+\; \beta y\, \mathbf{k} \times \mathbf{U_b} \;=\; -\nabla\Phi_{LN}, \tag{6}$$

where ϵ_D is now the mechanical dissipation (or mixing) rate *throughout the boundary layer* and $\mathbf{U}_b$ the mass-weighted average boundary layer perturbation transport,

$$\mathbf{U}_b \equiv \frac{1}{Z_T\ \rho(Z_T)}\int_0^{Z_T} \rho U dz \;\simeq\; \frac{1}{H_b\ \rho(H_b)}\int_0^{H_b} \rho\ U\ \ dz, \tag{7}$$

The top of the boundary layer (taken by LN to be the 700 mb surface) is at $Z_T = H_b + \overline{h_b} + h$, where $\overline{h_b}$ is the zonal mean deviation of boundary layer top from the reference height H_b, and h the remaining "eddy" perturbation in boundary layer height (see Fig. 1b). [Note, $|h|/H_b$, $\left|h/\overline{h_b}\right|$, and $\left|\overline{h_b}\right|/H_b$ are taken small.]

The perturbation (eddy) pressure gradient averaged over the depth of the boundary layer from the surface ($z = 0$) to the Z_T is written (see LN, Eq. 6c and 7c)

$$\frac{1}{Z_T\ \rho(Z_T)}\int_0^{Z_T} \nabla P dz \simeq \nabla\Phi_{LN}, \tag{8}$$

where

$$\Phi_{LN} \equiv g\left(1 + \alpha H_b/\overline{T}_s\right) h - \Gamma_{LN}\ T_v{}'(x,y), \tag{9}$$

and $\Gamma_{LN} = \frac{gH_b}{2\overline{T}_s}(1 - 2\gamma/3)$ and g is the gravitational acceleration. The two contributions to the perturbation pressure are the changing thickness of the boundary layer due to fluctuations at the top (h) and changes in the density of the air in the boundary layer due to thermal perturbations, $T_v{}'(x,y)$.

Lindzen and Nigam argued that in regions of deep convection the surface mass convergence would be rapidly vented by convection. Therefore, according to (9), the full effects of the hydrostatically induced pressure gradient would not be realized.

They incorporated this 'back pressure' effect through the vertically averaged continuity equation and further assumed the boundary layer could relax back to equilibrium in time ${\epsilon_T}^{-1}$:

$$-H_b \, \nabla \cdot \mathbf{U}_b \;\; \simeq \;\; w(Z_T) \;=\; \epsilon_T h \; . \tag{10}$$

In the steady state, the velocity at the top of the boundary layer is the entrainment velocity, taken proportional to the boundary layer perturbation height (mass convergence)[3].

Neelin (1989) noted the mathematical similarity of the Gill model (Eqs 1 and 2) to the LN model by rewriting equation (9) using (10) whereby the LN model equations become:

$$\epsilon_D \mathbf{U}_b \, + \, \beta y \, \mathbf{k} \, \mathbf{x} \, \mathbf{U}_b \;\; = \;\; -\nabla \Phi_{LN}, \tag{6}$$

$$\epsilon_D \Phi_{LN} \;\; + \;\; {C_{LN}}^2 \, \nabla \cdot \mathbf{U}_b \;\; = \;\; -\epsilon_D \;\; \Gamma_{LN} \;\; T_v'(x,y), \tag{11}$$

where

$$ {C_{LN}}^2 \;=\; gH_b \left(1 + \frac{\alpha H_b}{\overline{T}_s}\right) \frac{\epsilon_D}{\epsilon_T} \;\simeq\; gH_b \frac{\epsilon_D}{\epsilon_T} \;\equiv\; {C_B}^2 \frac{\epsilon_D}{\epsilon_T}. \tag{12}$$

The mathematical similarity between the LN (Eqs 6 and 11) model and the Gill model (Eqs 1 and 2) is now apparent, although the physics is fundamentally different. Further discussion is deferred until section (4).

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[3]Hence, radiation is assumed to have a time-scale that is longer than that of entrainment or mixing.

2.b. The Reduced Gravity Boundary Layer Model

We hereby adopt the physically intuitive geometry of a well mixed boundary layer of constant potential virtual temperature:

$$\Theta_v(x,y) = \Theta_o + \overline{\Theta}_v(x,y) + \Theta_v{}'(x,y) \quad \text{for} \quad 0 < z < Z_T, \tag{13}$$

where Θ_o is a reference temperature, and $\overline{\Theta}_v$ and $\Theta_v{}'(x,y)$ are the basic state and perturbation virtual potential temperature in the boundary layer, respectively. The top of the boundary layer will be of variable depth:

$$Z_T = H_b + \overline{h_b} + h \tag{14}$$

and capped by an unperturbed constant inversion of strength $\Delta\Theta_v$ (see Fig. 1c). The model is shown schematically in Fig. 1c. The notation in Eq (14) is as in the LN model except the top of the boundary layer (at $z = Z_T$ is now defined by the level at which the inversion is found rather than the level of a constant pressure surface. The boundary layer includes the mixed layer and, when present, the stratocumulus/cumulus cloud layer. There is abundant observational evidence for this model geometry in the tropical and subtropical marine environment (see, for example, Riehl (1954), Riehl (1979), Augstein et al. (1974), Stage and Businger (1981), Nicholls and Leighton (1986), including the regions nearby the intertropical convergence zones (Bunker (1971), Ramage et al. (1981)).

Following LN, we will assume that the boundary layer flow is driven directly by hydrostatically induced pressure gradients that are associated with the perturbations in the virtual potential temperature perturbations, $\Theta_v{}'(x,y)$ which are confined to be within the well-mixed boundary layer. Inasmuch, the role of the free atmosphere (above Z_T) is to *reduce gravity* and absorb mass and heat from the boundary layer *without* producing significant vertically averaged flow throughout the troposphere *above* the boundary layer. [Significant baroclinic flow above the boundary layer can result from deep convection, but we assume that the integrated effects of this flow at the top of the boundary layer are small compared to the boundary layer processes.][4]

We obtain, by a mass-weighted average of the momentum equation for the boundary layer, the equation for the surface flow:

$$\epsilon_D \mathbf{U}_b \; + \; \beta y \; \mathbf{k} \; x \; \mathbf{U}_b \;\; = \;\; -\nabla \Phi_{RG}, \tag{15}$$

where $\mathbf{U}_b$ is defined in Eq (7) and Φ_{RG} is the perturbation pressure for the reduced gravity system:

$$\Phi_{RG} \equiv g'h - \Gamma \; \Theta_v{}'(x,y) \; , \tag{16}$$

where $g' \equiv g\frac{\Delta\Theta_v}{\Theta_o}$ is the value of reduced gravity and $\Gamma = gH_b/2\Theta_o$ (terms of O($|h|/H_b$, $\frac{|\Theta_v{}'(x,y)|}{\Delta\Theta_v}$) have been neglected in Eq 16). Following LN, we will assume that, in the steady state, the rate of relaxation of the boundary layer perturbations will depend on the mass convergence in the boundary layer. The relaxation rate, however, will now depend on whether or not there is enough moisture to support convection. When convection (above the boundary layer) is diagnosed, the venting of the boundary layer will be taken as the time scale ${\epsilon_M}^{-1}$ associated with the life cycle of deep convection[5]. In the absence of deep convection, the mixing out of the boundary layer perturbations will be at the slower entrainment rate, ϵ_D, observed to be O(1/(1-2 days)) (see section (2.b)). Hence, the equations for the reduced gravity model of the tropical boundary layer flow driven by hydrostatically induced pressure gradients is (c.f., Eqs 6, 11, 12):

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[4]The model formulation is familiar to oceanographers as a 1 1/2 layer model forced by buoyancy, rather than wind stress.

[5]Note that the physics involved in the parameter ϵ_M is identical to ϵ_T used by LN. We distinguish the two by the numerical values assigned to this process, which differ by a factor of 50 or so.

$$\epsilon_D \mathbf{U}_b \ + \ \beta y \ \mathbf{k} \ \mathbf{x} \ \mathbf{U}_b \ \ = \ \ -\nabla \Phi_{RG} \tag{15}$$

convecting:

$$\epsilon_D \ \Phi_{RG} \ + \ \frac{\epsilon_D}{\epsilon_M} C_{RG}{}^2 \ \nabla \cdot \mathbf{U}_b \ \ = \ \ -\epsilon_D \ \Gamma \ \Theta_v{}'(x,y) \tag{17}$$

not convecting:

$$\epsilon_D \Phi_{RG} \ + \ C_{RG}{}^2 \ \nabla \cdot \mathbf{U}_b \ \ = \ \ -\epsilon_D \ \Gamma \ \Theta_v{}'(x,y) \ , \tag{18}$$

where

$$C_{RG}{}^2 \equiv g' H_b \equiv C_B{}^2 \frac{\Delta \Theta_v}{\Theta_o} \ . \tag{19}$$

It will prove to be convenient to write (17 and 18) as one equation:

$$\epsilon_D \ \Phi_{RG} \ + \ C_{RG}{}^2 \ (1 - \beta') \ \nabla \cdot \mathbf{U}_b \ \ = \ \ -\epsilon_D \ \Gamma \ \Theta_v{}'(x,y) \ , \tag{20}$$

where

$$\beta' \equiv \left\{ \begin{array}{cc} \epsilon_D \epsilon_M{}^{-1} & \text{if convecting} \\ 0 & \text{otherwise} \end{array} \right\} . \tag{21}$$

The primary differences between the boundary layer model presented here and the original LN boundary layer model are as follows: (i) the effect of gravity is reduced from that in LN by formulating a two-layer system and (ii) the boundary layer relaxation rate depends on the convective activity *above* the boundary layer. The latter effect acts to stiffen the boundary layer where convection is not occurring, hence producing convergence and pressure perturbations of smaller amplitude and greater horizontal extent.

2.c. The Precipitation Criterion and Venting Rate

It remains for us to ascertain whether the boundary layer circulation is supporting convection. We stress that, in the LN and RG models, convective process may only act to restore the boundary layer flow: diabatic heating via precipitation does not originate from within the boundary layer. In the steady state, precipitation must occur if the evaporation rate exceeds the moisture divergence due to the flow field. Since evaporation is at the sea surface and most of the moisture is found within the boundary layer, a diagnostic check for precipitation is provided by the moisture budget (see, e.g., Weare, 1986):

$$P \equiv \rho_{air} C_E \, |\mathbf{u}| \, (q(SST) - q(air)) \, |_{z=0} - \int_0^{Z_T} \nabla \cdot (\rho_{air} \, q(air) \, \mathbf{u}) \, dz - \delta_m \, , \quad (22)$$

where the bulk formula for evaporation has been assumed and $P \geq 0$. In Eq (22), δ_m is the loss of moisture from the boundary layer to the free atmosphere by ubiquitous mixing (detrainment).

If the circulation is found to support convection ($P > 0$), rapid venting is assumed for the boundary layer Eq (17). However, if Eq (22) indicates a net moisture deficit in the boundary layer ($P < 0$), then there is no rain in the steady state and the restoration of the boundary layer circulation is by the relatively inefficient process of mixing across the inversion (entrainment): Eq (22) applies. In practice, the model equations 20, 21, and 22 are solved iteratively.

In the calculations we have done (discussed in section 3), we have estimated δ_m from the climatological monthly mean fields of evaporation, moisture convergence and precipitation. While there is some spatial variation in δ_m, we assumed a uniform constant value of 10^{-5} kg m^{-2} s^{-1}. The solutions are not particularly sensitive to the value of δ_m.

2.d. The Surface Winds

The boundary layer model presented in section 2.b is for the mass transport averaged over the depth of the boundary layer. Neelin (1989) showed that the momentum equations (15) are a good approximation for the integrated boundary layer transport in the GFDL GCM and, hence, provide a reasonable estimate of the moisture convergence in the boundary layer. Nonetheless, the holy grail for investigators that use the steady (slave) atmospheric models is frequently the surface wind velocity (wind stress). Using the results of Deser (1993), the solution for the pressure from the reduce boundary layer model can be used to give a reasonable solution for the surface winds. Specifically, Deser showed that by using an asymmetric Rayleigh damping in Eq (15), a good estimate of the surface winds are given by the observed pressure gradients. Specifically, Deser found that a good estimate of the surface winds is found by replacing in Eq (15) $\epsilon_D \mathbf{U}_b$ with

$$\epsilon_D \mathbf{U}_b => {\epsilon_D}^x \, U \, \mathbf{i} \; + \; {\epsilon_D}^y \, V \, \mathbf{j}$$

where ${\epsilon_D}^y/{\epsilon_D}^x \;\approx\; 2-3$. Deser argued that the asymmetric Rayleigh damping represents the vertical structure within the boundary layer that is due to Ekman-like effects. We have incorporated the scheme of Deser (1993) into our reduce gravity boundary model to output the surface wind fields as well as the boundary layer mean winds.

3. Critique of Simple Tropical Atmosphere Models

A large body of literature has developed concerning large-scale atmosphere-ocean instability which is based on the atmospheric model formulated by Gill (1980) (see, e.g., Philander et al. (1984), Rennick (1983), Gill (1985), Hirst (1986)). This model atmosphere formulation is also utilized in most studies that have lead to the formulation of the potential mechanisms for the ENSO phenomenon and interannual variability in the tropics (see, e.g., Anderson and McCreary (1985), Zebiak and Cane (1987), Schopf and Suarez (1988), Battisti (1988), Battisti and Hirst (1989), Wakata and Sarachik (1991)). Yet the cumulative problems with the Gill model, noted in section (1) and summarized in Table 1, suggest this framework must be considered to have only qualitative value (see also, Seager, 1991). **Table 1.**

The formulation of the boundary layer model of LN, on the other hand, is somewhat more consistent with the observations. The same mathematical equations are realized in both models (c.f., Eqns 1 with 6, and 2 with 11), except the boundary layer model yields the flow averaged in the surface layer rather than the baroclinic tropospheric flow field. Neelin (1988) has presented calculations that indicate the momentum equation (6) forced by the 700mb vertical velocity from a GCM is indeed a good approximation to the flow over the ocean in the lowest 300 mb of a General Circulation Model with damping times $\epsilon_D^{-1} \simeq 2$ days. Zebiak (1990) has presented similar conclusions based on an analysis of observed flow and pressure fields. This suggests that linear dynamics may be sufficient for the boundary layer flow. In Neelin's (1988) calculation, the forcing at the top of the boundary layer (effectively the Φ_{LN} term in Eq. (6)) was prescribed from the GCM calculations. In general, the perturbation height will be due to heating of the boundary layer from surface fluxes, turbulent entrainment of dry air from above into the boundary layer, and detrainment from the top of the boundary layer; the later being sensitive to the presence of convection aloft.

The heating of the Gill model by Zebiak's (1986) iterative forcing scheme (Eq. 3) includes a convective heating term that is proportional to the SST anomaly, and a term that depends on the total (mass) convergence in the boundary layer. Zebiak's heating scheme can be approximated as follows (see Zebiak, 1985):

$$\underset{\text{total heating}}{Q_o} = \underset{(A)}{\Lambda\, T'} - \underset{(B),}{\beta_Z {c_a}^2 \nabla \cdot \mathbf{u}} \tag{23}$$

where

$$\beta_Z = \begin{cases} 3/4 & \text{convecting} \\ 0 & \text{not convecting.} \end{cases}$$

The term (A) is the evaporatively induced convective heating anomaly, while (B) is a CISK-like feedback term. Zebiak noted the magnitude of the phase speed (${c_a}^2$, ${C_{LN}}^2$) in the Gill and LN model is very similar, and the forcing functions (convective $\Lambda T'$, evaporative $\epsilon_D\ \Gamma\ T_v'(x,y)$) have similar dependence on SST anomalies and amplitude ($T_v'(x,y) \simeq$ SST′). Hence, Zebiak (1990) found that the Gill model solutions, forced with only the evaporative term (A), were comparable to the boundary layer model of LN, and neither looked like the observations. He further demonstrated that including the CISK term (B) in the heating of the Gill model yields surface circulation and pressure distributions more like the observations.

The good simulation of the surface flow field provided by the physically flawed "Gill model" (see Table 1 and section 1), when forced by Zebiak's (1986) heating scheme, can now be understood by contrasting the set of equations for the Gill model (with heating via Eq (23)) with the equations for the reduced-gravity boundary layer model presented in section (2.b):

Gill/Zebiak Model

$$\underline{\epsilon_D\, \mathbf{u}} \;-\; \beta y\, \mathbf{k} \times \mathbf{u} \;=\; -\nabla\Phi \qquad (1)$$

$$\underline{\epsilon_D\, \Phi} \;+\; {c_a}^2\, \nabla\cdot\mathbf{u} \;=\; -\underline{\Lambda\, T'} \;+\; \beta_Z\, {c_a}^2\, \nabla\cdot\mathbf{u}, \qquad (24)$$

Boundary Layer Model

$$\epsilon_D \mathbf{U}_b \;+\; \beta y\, \mathbf{k} \times \mathbf{U}_b \;=\; -\nabla\Phi_{RG}, \qquad (15)$$

$$\epsilon_D \, \Phi_{RG} \; + \; C_{RG}{}^2 \, \nabla \cdot \mathbf{U}_b \;\; = \;\; -\epsilon_D \, \Gamma \, \Theta_v{}'(x,y) \; + \; \beta' \, C_{RG}{}^2 \, \nabla \cdot \mathbf{U}_b. \qquad (25)$$

[Eqn (24) is from Eq (1) and (23); Eq (25) is Eq (20) rewritten.] The mathematical formulation of both models is identical: there are minor differences on how convection (β_Z; β') is diagnosed. In addition, the values of the key parameter combinations are similar:

Gill Model

$$\begin{array}{c} c_a = 60 \text{ ms}^{-1} \\ \Lambda = 2.5 \; x \; 10^{-3}\text{m}^2\text{s}^{-3}\text{K}^{-1} \end{array} \; => \nabla \cdot \mathbf{u} \simeq \frac{\Lambda}{c_a{}^2} = \; 7 \; x \; 10^{-7}\text{K}^{-1}\text{s}^{-1}$$

Reduced Gravity Boundary Layer Model

$$\begin{array}{c} C_{RG} = 17.9 \text{ ms}^{-}1 \\ \epsilon_D\Gamma = 6.4 \; x \; 10^{-4}\text{m}^2\text{s}^{-3}\text{K}^{-1} \end{array} \; => \nabla \cdot \mathbf{U}_b \simeq \frac{\epsilon_D\Gamma}{C_{RG}{}^2} = \; 1.95 \; x \; 10^{-6}\text{K}^{-1}\text{s}^{-1}$$

Hence, it is to be expected that the two models yield similar answers. However, we noted in section 1 that some fundamental assumptions made in the Gill formulation are inconsistent with the observed atmospheric flow (e.g., the inherent nonlinearity in the upper tropical tropospheric vorticity balance and, in the eastern and central equatorial Pacific, the decoupling of the surface and 850 mb flow fields). Additionally, some of the values required for the parameters in the Gill model are unrealistic: these terms are underlined in equations (1) and (24) above and summarized with the observed values in Table 1.

Turning now to the LN model, we note that the extraordinary venting time ϵ_T, required therein for realistic flow fields, is essentially compensated for in the present model by the hitherto neglected reduced gravity effect (see Table 1). Indeed, we expect a similar response of the two boundary layer models *in convective regions*, based on the comparison of the thermodynamic equations (11) and (20) (the momentum equation in the two models is identical). The only difference between the models is in the 'effective' phase speeds:

$$\frac{C_{LN}{}^2}{C_{RG}{}^2} = \frac{\epsilon_D/\epsilon_T}{(1-\beta')\Delta\Theta_v/\Theta_o} = O(1). \tag{26}$$

We have performed numerous experiments with the RG model, using a range of specified SST anomaly patterns for the tropical Pacific. The SST anomalies force of the RG boundary layer model by inducing anomalies in the "virtual sensible heat flux", being the total density effect of sensible heat and evaporative fluxes in the absence of condensation. When forced by the SST anomalies from, for example, a typical warm ENSO event, the RG model reproduces the observed anomalies in the boundary layer transport, the surface wind, and (diagnostically) the regions of anomalous precipitation with fidelity similar to that of the previously successful work using the Gill/Zebiak and LN models. The fields are therefore not shown. Some improvement in the flow simulations using the reduced gravity boundary layer model is due to the proper identification and treatment of the regions of *nonconvection* through the diagnostic moisture budget calculation. For example, the RG model (Eqs 15, 20-22) captures the southward displacement of the ITCZ in the eastern Pacific (and the assoicated meridional wind anomalies) during the mature phase of the Rasumsson and Carpenter composite ENSO. However, the general similarity of the solutions with those previously calculated is as expected given the similar form and coefficient values of the different models. The point here is that the form and values may be obtained, via the reduced gravity boundary layer model, without recourse to physically doubtful values for the forcing or damping coefficients.

Wang and Li (1993) also pointed out the deficiencies in the Gill/Zebiak and LN model formulations, and showed that a combined approach involving a baroclinic mode free atmosphere interacting with a planetary boundary layer was able to provide a broadly realistic surface wind pattern using realistic values of model parameters connected to boundary layer venting and thermal damping. However, he had to retain an extremely strong Rayleigh friction (approx. 1 day^{-1}) for the baroclinic mode free atmosphere in order to obtain realistic solutions. Moreover, he took the top of the boundary layer to be a fixed isobaric surface, and found that his solution was very sensitive to the assumed depth of this boundary layer. When the boundary layer is instead treated as a variable depth layer subject to reduced gravity, we find that there is no need to include an explicit baroclinic free atmosphere component with it's problematic damping coefficients.

4. Concluding Remarks

We present a new model of the steady tropical surface circulation that occurs as a result of spatial variation in the SST pattern. The model is similar to the Lindzen and Nigam (L/N) boundary layer model, driven by sensible heat and evaporative flux anomalies, only here the boundary layer is assumed well mixed and capped by an inversion. Hence, the model reduces to a two-layer, reduced gravity system. Furthermore, the rate of exchange of mass across the boundary layer/free atmosphere interface is dependent on the moisture budget in the boundary layer. When convection is diagnosed to occur, detrainment operates on the time scale associated with the life-cycle of deep convection (approximately eight hours). Otherwise, the detrainment time scale is assumed to be associated with the mixing out of the stable tropical boundary layer (approximately one day). The model provides a diagnostic estimate of the anomalies in precipitation, though the latent heat released is above the boundary layer and is assumed to drive a circulation that does not impact the boundary layer.

We discuss the inconsistencies between the formulations of both the Gill/Zebiak (G/Z) and Lindzen/Nigam model and the values of several of the parameters that are required for these models to achieve realistic solutions (circulations). Then, the new reduced gravity boundary model equations are re-written in the form of the G/Z and

L/N models. Using realistic values for the parameters in the new model configuration, we show that the constants combine in the re-written equations to produce the unphysical constants in the G/Z and L/N models; hence, the reason for the apparent success of these models.

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Postscript This paper was written in 1990. We had intended to augment the work presented here with a section on the role of the Andes (which eliminates some of the 'excessive easterly' problems in the Gill model) and with extensive calculations. Though we lost interest in the work and did not complete these sections, the work that we did complete is requested (and cited) regularly enough that we feel obligated to publish the completed portions of the work. For the many people who have a copy of the original manuscript, the only changes in the published version of the paper are (i) a short discussion in section (3) of the model by Wang and Li (1993), which had yet to be developed when we wrote the original manuscript, and (ii) an explicit discussion in section (2.d) of the surface winds as a function of the integrated mass transport in the boundary layer, which is inspired by the work of Deser (1993).

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Figure Caption

Figure 1. Schematics of the steady state atmosphere models discussed in this paper. (a) The Gill/Zebiak model (section 1). (b) The Lindzen/Nigam model (section 2.a). (c) The Reduced Gravity Boundary Layer model (sections 2.b and 2.c). The Gill model is a model of the circulation anomalies in the whole troposphere that are forced by anomalies in the diabatic heating (latent heat released) Q_o in deep convective clouds; Q_o has contributions from local evaporation Q_{evap} and vertically integrated moisture convergence Q_{conv}. It is commonly assumed in the Gill-type models that the winds extend from the base of the (elevated) heating to the ground.

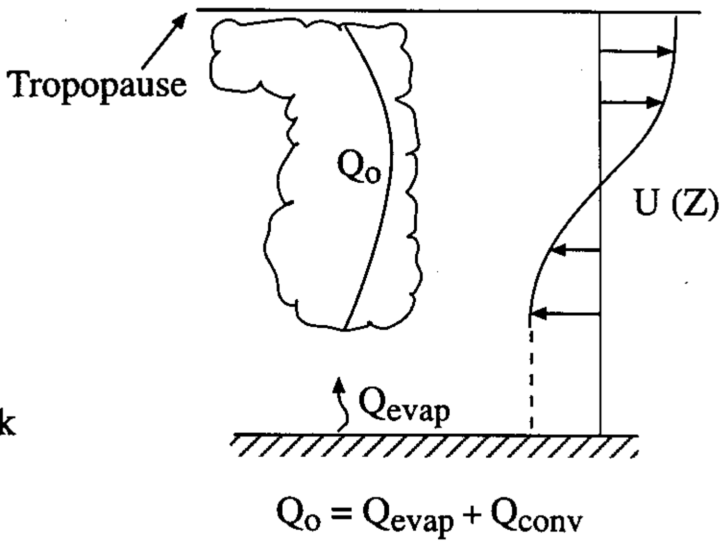
The Lindzen/Nigam and RG models are for the boundary layer response to anomalies in the "virtual sensible heat flux" Q_{vs} that result from changes in SST (SST'). Q_{vs} is the total density effect of sensible heat and evaporative fluxes in the absence of condensation. The Lindzen/Nigam model assumes a linear vertical profile in basic state ($\bar{T}$) and anomalous (T') temperature; air temperature anomalies at surface are equal to SST', and linearly decay to zero at height $z = Z_T$ where $P(Z_T) \equiv 700$ mb. In contrast, the RG boundary layer model assumes that the virtual sensible heat flux gives rise to virtual potential temperature anomalies that are well mixed in the boundary layer. The top of the boundary layer (at $z = Z_T$) is determined by hydrostatic effects and mixing across the boundary layer; the boundary layer is capped by a potential temperature jump of $\Delta\theta_v$. In the Lindzen/Nigam (RG) models, the climatological mean height of the 700 mb surface (inversion) is at $z = Z_T = H_b + \bar{h}_b$.

Table Caption

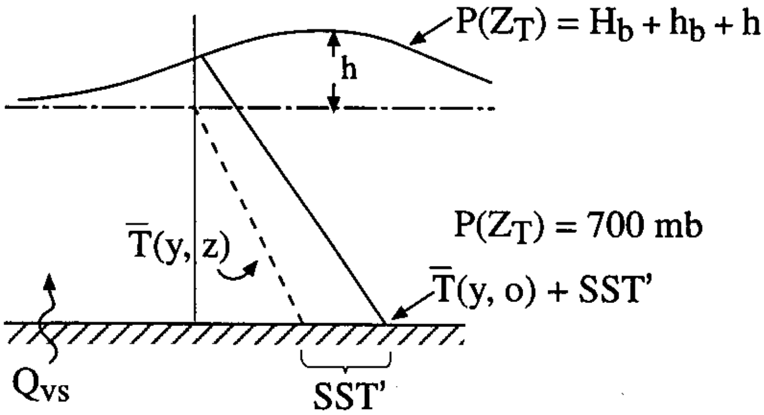
Table 1. Values for the constants used in various steady state tropical atmosphere models. The values that are in bold are qualitatively inconsistent with the observations. The * indicates a wave speed given by Eq. (12).

Table 1: Values for the constants used in various steady state tropical atmosphere models. The values that are in bold are qualitatively inconsistent with the observations. The * indicates a wave speed given by Eq. (12).

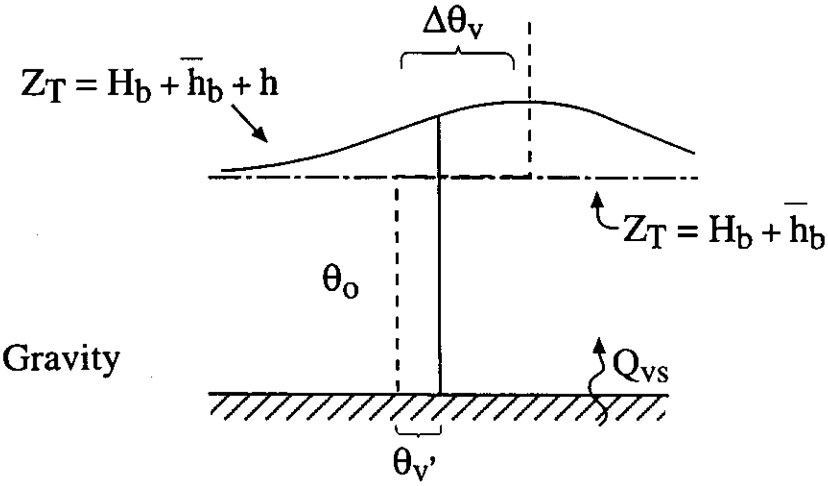
Model Type	Model Source	Mechanical Damping $\epsilon_D{}^{-1}$ (days)	Venting or Thermal Damping $\epsilon_T{}^{-1}$	Wave Speed m/s	Typical Forcing (Wm^{-2})
Convective	Gill/Zebiak	**1-2**	**1-2 days**	60	**150**
	Observations	$\approx$ 15	$\approx$ 15 days	$\approx$ 40	25-40 (evaporative)
Boundary Layer	Lindzen-Nigam	$\approx$ 1-2	$\approx$**30 minutes**	$\approx$ 16*	15 (virtual sensible)
	Reduce Gravity (section (2b)	$\approx$ 1-2	$\approx$ 8-12 hours	$\approx$ 18	15 (virtual sensible)
	Observations	$\approx$ 1-2	$\approx$ 8-12 hours	$\approx$ 18	15 (virtual sensible)



(a) Gill / Zebiak



(b) Lindzen / Nigam



(c) Reduce Gravity