Large-scale and evaporation-wind feedbacks in a box model of the tropical climate

G. Bellon and H. Le Treut

Laboratoire de Météorologie Dynamique du CNRS, IPSL, Université Pierre et Marie Curie, Paris, France

M. Ghil

Department of Atmospheric Sciences and Institute of Geophysics and Planetary Physics, University of California, Los Angeles, USA, and Département Terre-Atmosphère-Océan and Laboratoire de Météorologie Dynamique du CNRS, Ecole Normale Supérieure, Paris, France.

Abstract. A four-box model of the tropical climate is used to assess its sensitivity to changes in the relative area of the moist and dry regions and to radiative perturbations. The feedbacks modulating this sensitivity are analyzed, particularly the dynamical feedbacks associated with the surface fluxes. The link between the large-scale circulation and these fluxes is found to play a crucial role in the tropics' sensitivity to radiative perturbations.

1. Introduction

The observed tropical sea-surface temperatures (SSTs) exhibit a maximum around 30^{o}C; this maximum appears to be robust on various timescales, from intraseasonal to millennial. The robustness of this apparent threshold seems to contradict the fact that the main radiative feedback acting on the SST, which is due to water vapor, is generally considered to destabilize the system. Identifying the stabilizing feedback(s) that help(s) maintain this threshold is essential in order to understand how the tropical climate reacts to an external perturbation. Many hypotheses have been proposed such as the drying of nonconvective zones (Sun and Lindzen, 1993), the increase or decrease of the area covered by cirrus clouds (Ramanathan and Collins, 1991; Lindzen et al., 2001), or the effects of the large-scale circulation on the local radiative-convective equilibrium (Wallace, 1992; Hartmann and Michelsen, 1993, hereafter HM93).

The identification and tentative validation of such atmospheric feedbacks has also been carried out extensively using general circulation models (GCMs; Cess et al., 1996). In most cases, however, sensitivity analyses of tropical climate have been conducted in the conceptual framework of one-dimensional column models; this was the case even when using a large number of independent vertical columns, as in the traditional diagnostics of GCM results in terms of radiative perturbations.

The low latitudes are characterized by the homogeneity of the temperature profiles and by the bimodality of the humidity profiles in the free troposphere. Pierrehumbert (1995, hereafter P95) used a simple two-box model to show that the coupling between the two humidity profile modes in the Tropics and the large-scale circulation helps maintain the observed tropical climate: the circulation transports the excess of energy from the convecting box, which is in a state of runaway greenhouse, to the subsiding box, where it is radiated away to space. P95's minimal description of the tropical energetics clearly demonstrates that the tropical climate is not determined locally, but globally, and that the interactions between moist and dry regions are an essential part of its stability as well as its variability. Various factors have been shown to affect this radiative-convective two-column equilibrium: low-level clouds (Miller, 1997), ocean transport (Clement and Seager, 1999), as well as boundary layer properties and the hydrological cycle (Larson et al., 1999, hereafter L99).

These box models have been shown to be quite sensitive to the relative area of the moist and dry regions. The present work analyzes further the feedbacks associated with this sensitivity to show how they modulate the response of the tropical temperature to a radiative perturbation. Furthermore, we investigate the influence of the surface-wind parameterization: the link between the surface fluxes and the large-scale circulation is varied systematically in order to shed some light on the underlying feedbacks.

2. Model Formulation

2.1. Description

The model we use here (Fig. 1) is analogous to that of L99. One column represents the deeply convecting Tropics, with the underlying warm pool, while the other one stands for the subsiding regions. The free troposphere in the subsiding column is thermodynamically uncoupled from the boundary layer by the inversion layer, but it is coupled to the convective column: the temperature profile in the free troposphere is approximatively homogeneous throughout the Tropics, while the free-tropospheric humidity in the subsiding column mainly originates from the convective column (Sun and Lindzen, 1993). Hence, we define our four model boxes as follows: the boundary layer of the subsiding column, the rest of the atmosphere, and the warm and cold pools. The model emphasizes the role of the boundary layer in the nonconvecting regions as a reservoir of moisture for convection. The model's dry free troposphere, whose thermodynamic structure is linked to that of the convective regions, controls the radiative input into the boundary layer, while its radiative cooling controls the strength of the large-scale circulation.

We use log-pressure coordinates in the vertical: $\zeta = -H \log(p/p_0)$, where H = 8 km and p_0 = 1 atm is the surface pressure. Besides the parameterizations of the surface

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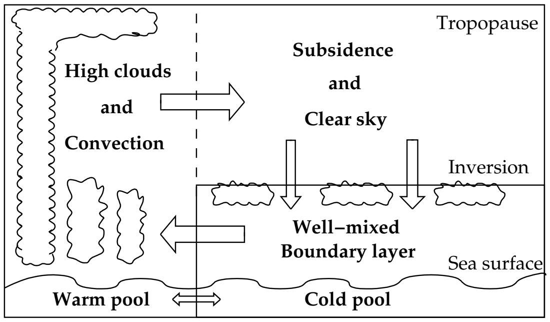

Figure 1. Schematic model diagram.

fluxes and radiation that are explained in the next subsections, our model is very similar to that of L99. Its main features are outlined below, while noting several simplifications that were introduced in order to make our model more flexible as a sensitivity-study tool:
(i) The moist adiabat giving the temperature profile in both the convective column and the dry free troposphere is approximated by a fixed lapse rate $\Gamma = 0.006$ Km^{-1}, as in P95, above a well-mixed sub-cloud layer of given thickness $\zeta_1 = 700$ m.
(ii) The relative humidity is set to 75% in the convective column (P95).
(iii) The trade-wind boundary layer is assumed to be well mixed, and the altitude p_{inv} of the inversion, in pressure coordinates, is parameterized following Sarachik (1985):

$$\frac{dp_{\mathrm{inv}}}{dt} = \alpha_s g \frac{LE}{C_P \Delta T_v^{inv}} \approx \alpha_s g \frac{LE}{C_P \Delta T^{inv}}; \qquad (1)$$

here g is the acceleration of gravity, C_P the heat capacity of air, LE is the latent heat flux at the surface, ΔT^{inv} (resp. ΔT_v^{inv}) is the temperature (resp. virtual temperature) jump at the inversion, and α_s is a constant, taken to be 0.3.
(iv) The profile of the mass flux between the two columns in the free troposphere is chosen to decrease linearly from the inversion to the tropopause ($\zeta_{\mathrm{top}} = 15$km), where the flux is zero. This profile mimics the detrainement of convective anvils and produces a realistic free-tropospheric humidity profile in the dry column. The total mass flux between the columns is diagnozed considering the balance between heating due to subsidence and radiative cooling in the dry free troposphere.

The balances of moist static energy and total water between the four boxes allow us to compute the precipitation rate in the warm pool, the air and sea surface temperatures, the surface humidity in the dry boundary layer, and the subsidence at the inversion level in equilibrium.

2.2. Radiation

We implemented a very simple radiation code: For the incoming solar radiation, the clear-sky diffusivity is neglected. For the outgoing infrared (IR) radiation, the surface is considered as a black body and the atmosphere as a grey body. The absorption a of the latter is parameterized as a function of humidity:

$$a = \rho(\alpha q + \beta); \qquad (2)$$

here ρ is the air density, q the specific humidity, α a proportionality coefficient, and β a parameter that represents the absorption of atmospheric trace constituents. The IR

Table 1. Values of the model parameters in the reference equilibrium

Parameter	Value
Area of the convective column A_c	33 %
Proportionality constant λ in Eq. (5)	0.35
Export to the extratropics	40 Wm^{-2}
Insolation C_s	350 Wm^{-2}
Surface albedo	0.06
α	5.0 10^{-2} kg^{-1}m^2
β	1.5 10^{-4} kg^{-1}m^2

contribution is integrated for each box following Fouquart (1988).

The radiative effect of clouds is fixed: the low clouds below the inversion are represented by an additional albedo of 0.1. The convective clouds reflect the incoming solar radiation and also add to the greenhouse effect. These clouds'net forcing at the top of the atmosphere is still a controversial issue (Lindzen et al., 2001; Hartmann et al., 2001) but its magnitude seems limited. We neglect this net forcing, as well as the redistribution of the radiative cooling between the ocean and atmosphere due to these clouds.

2.3. Surface fluxes

The surfaces fluxes of sensible heat F_S and latent heat LE are parameterized using bulk formulas:

$$F_S = \rho_s C_d V_s C_P [T - T(0)], \qquad (3)$$

$$LE = \rho_s C_d V_s L [q_*(T) - q(0)], \qquad (4)$$

where ρ_s is the density of the surface air, V_s the surface wind, T the SST, $T(0)$ the surface-air temperature, L the latent heat of vaporization of water, $q(0)$ the surface air's specific humidity, C_d a drag coefficient equal to 0.0013, and q_* the saturated specific humidity. The total surface heat flux is given by $F = LE + F_S$ and is dominated by the latent heat flux LE (about 90% of the total in the Tropics).

In reality, the surface wind V_s depends on the large-scale circulation and this dependence introduces a wind-induced surface heat exchange (WISHE) effect (Emanuel et al., 1994): the surface fluxes increase with the intensity of the large-scale circulation, an effect that warms the entire atmosphere. This dependence is complex and difficult to parameterize at our model's level of simplicity: the large-scale circulation is related to the divergence of the wind in the lower troposphere, but the nondivergent component of the surface wind and the small-scale fluctuations cannot be explicitely modeled. We thus choose to arbitrarily link the surface wind to the large-scale circulation as follows:

$$V_s = (1 - \lambda) V_{s_0} + \lambda \eta m. \qquad (5)$$

Here V_{s_0} is a constant reference wind, equal to 6.5 ms^{-1}, and η is a proportionality coefficient set to $2.25.10^3$ kg^{-1}m^3. The total mass-flux m exchanged by the boxes is given by $m = (1 - A_c)\omega_{\mathrm{inv}}/g$, where A_c is the relative area of the convective column, and ω_{inv} the vertical speed at the inversion; λ is a parameter describing the dependence of the surface wind on the large-scale circulation: if λ=0, the surface wind is independent of the large-scale circulation, while for λ=1 the surface wind is proportional to the large-scale circulation.

3. Equibrium States and Sensitivity

3.1. Reference equilibrium and methodology

The model described in Section 2 is integrated in time until it reaches an equilibrium. Table 1 lists the values of the

Table 2. Results for the reference equilibrium

Variable	Warm pool	Cold pool
T, the SST	303.0 K	300.0 K
$T(0)$, the surface-air temperature	302.8 K	297.7 K
$q(0)$, the surface specific humidity	19.1 g/kg	15.2 g/kg
$\omega_{\rm inv}$, the subsidence rate	—	38.4 mb/day
$p_{\rm inv}$, the altitude of the inversion	—	846 mb
Precipitation rate	505 cm/yr	—
F, the total surface heat flux	133 $\rm Wm^{-2}$	150 $\rm Wm^{-2}$
m, the large-scale overturning	3.10^{-3} kg m^{-2} s^{-1}	

parameters used to obtain this reference equilibrium. The atmospheric and oceanic exports to the extratropics are assumed to be equal and fixed. The results are given in Table 2: they are close to the climatological values as well as to the results of other models (L99, and references therein). **[Tables 1 and 2 near here, please]**

3.2. Sensitivity to radiative and areal perturbations

The sensitivity of this type of model to the warm pool's relative area A_c is significant, according to L99; it is interesting, therefore, to compare the orders of magnitude of the model sensitivities with respect to A_c and to radiative perturbations. To do so, we study three cases: a 10%-increase of the convective column area (A_c=43%, Case A_c+10), a doubling of carbon dioxide concentration in the atmosphere ($\beta=1.7.10^{-4}$ kg^{-1}m^2, Case 2*CO_2) and an increase of the insolation (C_s=355Wm^{-2}); the latter Case C_s+5 – corresponding roughly to the radiative forcing at the tropopause caused by a CO_2 doubling. **[Table 3 near here, please]**

As expected, an SST increase is obtained in all three cases (see Table 3). In our model, a doubling of the CO_2 concentration gives an increase in SST of more than 2^{o}C, 70% larger than a 5-Wm^{-2} perturbation in the solar constant C_s; the change caused by a 10% increase of the convective column's area is of the same order of magnitude as the latter. Actually, the strength m of the overturning circulation increases with a positive radiative perturbation, because the radiative cooling of the dry free troposphere, and thus $\omega_{\rm inv}$, increase; consequently, the transport of energy from the convective zone to the subsiding one, where it can be radiated to space, increases. This dynamical feedback reduces the warming perturbation. Inversely, in the Case A_c+10, the change in the area of the subsiding region compensates the change in the subsidence rate and the strength m of the overturning circulation decreases slightly. The warming is thus enhanced by the large-scale feedback and reaches the same order of magnitude as the warming due to a radiative perturbation. These feedbacks are illustrated in Fig. 2.

The figure shows that the constraints on the relative sizes of the convective and subsiding regions given by the dynamical feedbacks of the tropical atmosphere could either enhance or damp significantly the SST change caused by a

Table 3. Model sensitivity to areal or radiative perturbations. The columns represent change in T, the average SST; the SST difference ΔT between the warm and cold pool; the change of the subsidence rate $\omega_{\rm inv}$ at the inversion (in mb/day); the change of the overturning circulation m (in kgm^{-2}s^{-1}); and the change in the temperature difference $T - T(0)$ between the SST and the surface-air temperature.

Change in	T (K)	ΔT (K)	$\omega_{\rm inv}$	m	T–$T(0)$ (K)
A_c+10	+1.6	+0.52	+5.8	-3.10^{-5}	-0.73
2*CO_2	+2.2	+0.65	+9.6	$+8.10^{-4}$	-1.42
C_s+5	+1.3	+0.38	+5.0	$+4.10^{-4}$	-0.68

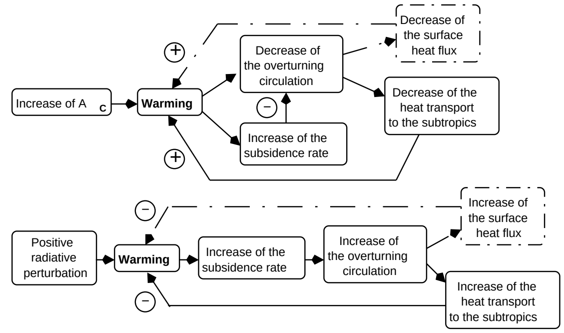

Figure 2. Feedback loops for an areal or radiative perturbation. Plus: positive feedback; minus: negative feedback. Dash-dotted: feedback added by the dependence of the surface wind on the large-scale circulation

radiative perturbation. The dynamics in the lower levels of the tropical atmosphere are at least partially controled by the surface temperature gradients (Lindzen and Nigam, 1987). The SST difference between the two boxes is increasing in all three cases (see Table 3). This effect is due to the bimodality of the radiative cooling: the contrast between the runaway greenhouse effect in the convective column and the transparency of the free troposphere in the subsiding column causes the differential warming.

The SST difference between the two regions is consistently correlated with a warming:

$$\frac{\partial \Delta T}{\partial T} \approx 0.3, \tag{6}$$

which shows that the low-level convergence would tend to increase with a warming. The competition between the change of the large-scale circulation imposed by the clear-sky thermodynamics and the enhancement of the low-level convergence is therefore a key process in determining the relative extent of the convective and subsiding regions.

3.3. Feedbacks associated with the surface flux

Because of the nonlinearity of the Clausius-Clapeyron equation and the dominance of the evaporation in the surface heat flux F, changes in F provide a strong negative feedback for a change in SST, if $T - T(0)$ and V_s remain constant (HM93). Our model suggests that the sea–air temperature difference $T - T(0)$ decreases with a warming (see the last column in Table 3). A direct consequence is that $T(0)$, and hence the entire atmosphere, warm more than the underlying SST. This enhanced atmospheric warming has two

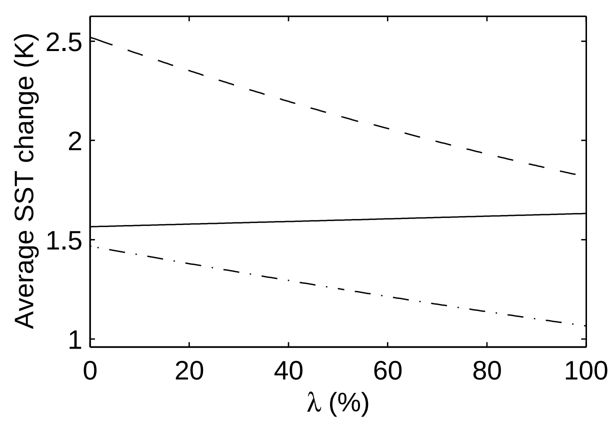

Figure 3. Sensitivity of the warming on the parameterization of the surface wind; Case A_c+10: solid line; Case 2*CO_2: dashed line; Case C_s+5: dash-dotted line.

effects: (i) the average outgoing IR flux at the top of the atmosphere increases, thus providing a negative feedback on the temperature of the whole system (i.e. the atmosphere acts as a buffer); and (ii) the surface fluxes and thus the corresponding negative feedback are reduced.

Furthermore, the large-scale circulation interacts with the surface fluxes by ventilating the boundary layer, but it also has an impact on the surface wind V_s (Emanuel et al., 1994; Bates, 1999). To study the latter influence in detail, we plot in Figure 3 the SST changes as a function of λ, the parameter that describes the dependence of V_s on m, for the three cases described in the previous section.

In our model, the wind-induced surface heat exchange (WISHE) introduces an additional feedback whose sign and magnitude depend directly on the sensitivity of the large-scale circulation to the perturbation: an increase of m enhances the surface heat flux F. Increasing F allows, in turn, a cooling of the global system because it transfers the heat from the ocean to the atmosphere, where it can be exported by the large-scale circulation to clear-sky regions and radiated to space. From the previous section, it follows that WISHE exerts a negative feedback for a radiative perturbation, while it has a slightly positive feedback for a change in A_c (see Fig. 2). WISHE can thus eliminate up to a quarter of the warming due to a radiative perturbation (see Fig. 3).

We can characterize the magnitude of the surface flux feedback by the surface heat flux increase normalized by the temperature change: $\partial F/\partial T$. The HM93 model takes into account the ventilation of the boundary layer by the large-scale circulation and gives a surface flux feedback of 7.9 $\mathrm{Wm^{-2}K^{-1}}$. Our model gives a slightly smaller feedback for λ=0, of 5.5 to 6.5 $\mathrm{Wm^{-2}K^{-1}}$. The difference between the two models is explained by the decrease of the air–sea temperature difference in ours, while this difference is fixed in HM93.

In our model, the surface flux feedback changes almost linearly with λ; it increases significantly for a radiative perturbation while it decreases slightly for a perturbation of the area:

$$\frac{\partial}{\partial \lambda}\left(\frac{\partial F}{\partial T}\right)^{rad} = 4\ \mathrm{Wm^{-2}K^{-1}}, \qquad (7)$$

$$\frac{\partial}{\partial \lambda}\left(\frac{\partial F}{\partial T}\right)^{A_c} = -0.5\ \mathrm{Wm^{-2}K^{-1}}; \qquad (8)$$

Consequently, the surface-wind dependence on the large-scale circulation increases the negative feedback associated with the surface flux and reduces substantially our model's sensitivity to radiative perturbations. Besides, this dependence increases the sensitivity to A_c relative to the sensitivity with respect to a radiative perturbation.

4. Concluding remarks

Our model's sensitivity to the relative areas occupied by the two types of radiative-convective and dynamical behavior — moist and ascending vs. dry and descending — in the tropical atmosphere is quite substantial. This sensitivity suggests the presence of an important and as-yet unexplored feedback in Earth's tropical climate, that could contribute to maintain the "lid" on tropical SSTs. But whether the relative area A_c of the convective region would actually increase or decrease as a response to global warming is not clear.

This state of affairs requires further investigation of the processes that control the relative area of the dry and moist regions. Lindzen et al. (2001) have argued that the cirrus-covered area — which can be taken as an approximation of our convective-box area — is regulated by cloud microphysics. On the planetary scale that our model emphasizes, dynamical processes must also play an important role in controlling A_c: recent papers by Kelly and Randall (2001) and Bretherton and Sobel (2002) are a first step in a further study of this role.

Our results show that the dependence of the surface wind on the large-scale circulation has an important effect on the sensitivity of the tropical system. This dependence reduces significantly the SST sensitivity to radiative perturbations by enhancing the evaporation feedback. More generally, the evaporation-wind feedback reinforces the large-scale circulation effects by injecting more heat into the atmosphere and allowing the atmospheric circulation to export more energy to the subtropical free troposphere, where this energy can be radiated to space.

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G. Bellon, LMD, Université Paris VI, Case 99, 4 place Jussieu, 75252 Paris Cedex 05, France(bellon@lmd.jussieu.fr)

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