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Surface wind, subcloud humidity and the stability of the tropical climate

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ABSTRACT

The dominant balance in the tropical ocean surface energy budget is between radiative heating and evaporative cooling. It has been proposed recently that a positive correlation between surface wind and ocean temperature could enhance the stabilising effect of evaporation. The issue is studied here using a single-column radiative-convective model of the tropical atmosphere. It is found that the proposed coupling of surface wind to temperature will indeed always increase the stability of the system, but the extent to which it does so depends in an essential way on the strength of the upper tropospheric water vapour feedback.

1. Introduction

Much work has been devoted in recent years to the identification of "thermostat" mechanisms for the tropical climate: that is, strong negative feedback mechanisms that tend to damp the response of the Tropics to temperature perturbations. A number of such mechanisms have been proposed, including adjustment of cirrus (Ramanathan and Collins, 1991) and stratus (Miller, 1997) clouds, upper tropospheric humidity (Sun and Lindzen, 1993) and ocean circulation (Clement et al., 1996). Recently, Bates (1999, hereafter B99) has proposed a stabilising mechanism based on angular momentum (AM) considerations. The argument runs briefly as follows. Climatologically, there is a net poleward flux of AM in the atmosphere, peaking at around 30° latitude in both hemispheres (Peixoto and Oort, 1992). To balance the AM flux convergence, a source (sink) of momentum is required in low (high) latitudes; this is provided by the climatological surface easterlies (westerlies) there. The atmospheric AM flux is due predominantly to transient eddies; to the extent that such eddies arise from baroclinic processes in the extratropics, it is plausible that the level of eddy activity (and hence the magnitude of the AM flux) will depend on the mean meridional temperature gradient. Thus, on the appropriate time scales, we expect a positive correlation between the meridional temperature gradient and the intensity of the surface zonal winds (we will refer to this as "AM coupling"). Evidence in favour of AM coupling is found in both observations (B99) and in GCM simulations (Bush and Philander, 1998; Alexeev and Bates, 1999). Now suppose that the tropical surface temperature is perturbed away from equilibrium (by a positive amount, say), while extratropical temperatures are held fixed. The tropical atmosphere will respond by warming up, thus increasing the meridional temperature gradient. By the argument above, surface winds will then also increase. Since sensible and latent heat fluxes are approximately proportional to surface wind, this will presumably lead to enhanced cooling of the surface, acting to bring

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the system back to equilibrium. We thus have a negative feedback loop.

In B99, this feedback mechanism was studied in the context of a simple, highly parameterised model atmosphere interacting with a mixed-layer ocean, and was indeed found to provide a substantial stabilising effect on SST. However, B99 made the crucial simplifying assumption that near-surface relative humidity remains constant in the face of changes in SST. This assumption, common to much previous work (for instance Newell and Dopplick, 1979; Hoffert et al., 1983; Hartmann and Michelsen, 1993), is based on the observation that near-surface relative humidity in the Tropics does indeed fluctuate by only a few percent despite large seasonal and interannual changes in SST. However, it is easily shown (Section 2) that a change of only a few percent in relative humidity can have an overwhelming impact on the evaporation rate. When the surface wind increases, we expect a tighter coupling across the surface layer and hence higher near-surface humidity. Conceivably, the increase in humidity could be strong enough that AM coupling could actually be a destabilising feedback.

Clearly then, it is important to asses the feedback mechanism proposed in B99 in a more realistic context where the constant relative humidity assumption is relaxed. That is the purpose of the present paper. We will continue to assume a positive correlation between meridional temperature gradient and surface wind, parameterising it as in B99, and employ the single-column radiative-convective model of Betts and Ridgway (1989, hereafter BR89) to determine the atmospheric temperature and humidity. The constraint of constant subcloud relative humidity is replaced by the more physically-based requirement that the atmosphere be in equilibrium at all times. This constraint is appropriate for time scales of a month or so, which are long compared to typical atmospheric equilibration time scales but short relative to the evolution of the ocean mixed-layer temperature.

In Section 2 the question of evaporative feedback is reviewed and a more precise formulation of the problem is given. The model atmosphere is described in Section 3 and its response to an SST perturbation is studied in Section 4. In Section 5 a simplified form of the model is presented which permits easier qualitative analysis of the processes determining subcloud humidity. Conclusions are drawn in Section 6.

2. Evaporation and the stability of the tropical SST

The evolution of the ocean mixed-layer temperature T_s is governed by

$$c\dot{T}_s = H = N_0 + F_l + F_s + F_o, \tag{1}$$

where c is the heat capacity (per unit surface area) of the ocean mixed layer, N_0 is the net surface radiative flux (short + longwave), and F_l and F_s are the surface latent and sensible heat fluxes and F_o is a lateral ocean transport term; H is the net energy budget of the ocean. Let us assume that the atmosphere and ocean are in equilibrium ($H = 0$) and perturb the ocean temperature by an amount δT_s. To first order, the perturbation will evolve according to

$$c\delta\dot{T}_s = \beta\delta T_s, \tag{2}$$

where (neglecting changes in ocean export F_0)

$$\beta = \frac{\partial H}{\partial T_s} = \frac{\partial N_0}{\partial T_s} + \frac{\partial F_s}{\partial T_s} + \frac{\partial F_l}{\partial T_s} \tag{3}$$

sets the rate at which the perturbation decays back to equilibrium, and thus characterises the stability of the system. In fact, to first order, β also controls the sensitivity of T_s to changes in an external parameter λ (such as insolation or atmospheric CO_2 concentration) through the relation

$$\delta T_s = -\beta^{-1}\frac{\partial H}{\partial \lambda}\delta\lambda$$

(Pierrehumbert, 1995). This identification of stability and sensitivity only works to first order, since at higher order cross terms appear in the above relation which will be absent in (2). Only the simpler issue of stability will be dealt with here.

If, as suggested by the geological record, the tropical SST is very stable, then we need to find mechanisms that give a large negative β. Let us concentrate on evaporation mechanisms, which come in through the latent heat flux term. To a good approximation (Hartmann and Michelsen, 1993) we may write

$$F_l \simeq \frac{L}{g}\omega_0(r-1)q_s,$$

where L is the latent heat of evaporation, r is the relative humidity of near-surface air, and q_s is the saturation specific humidity at the surface (note that F_l is taken as negative since it acts to cool the ocean). ω_0 is a surface flux parameter given by

$$\omega_0 = g\rho_a C_D |U|,$$

where ρ_a the surface air density, $C_D = 1.3 \times 10^{-3}$ is an empirical coefficient and $|U|$ is the magnitude of the surface wind. Thus

$$\frac{\partial F_l}{\partial T_s} \simeq F_l \left(\frac{1}{q_s}\frac{\partial q_s}{\partial T_s} + \frac{1}{\omega_0}\frac{\partial \omega_0}{\partial T_s} + \frac{1}{r-1}\frac{\partial r}{\partial T_s} \right). \quad (4)$$

The first term on the RHS depends only on the Clausius–Clapeyron relation, which states that saturation vapour pressure always increases with temperature. Thus this term always gives a negative, stabilising contribution to β (remember $F_l < 0$). For typical tropical conditions, this contribution is about -8 W m^{-2} K^{-1}. The results of B99 suggest that the ω_0 term in (4) (the "wind factor") will make a negative contribution of similar magnitude to that of the q_s term. Taken together, these two terms strongly stabilise the tropical SST.

The trouble lies in the third term on the RHS of (4) (the "relative humidity factor"). Assuming fixed r, this term is zero. However, a quick calculation (assuming $r = 80\%$) shows that a change in relative humidity of only 3% K^{-1} will give a contribution of magnitude greater than the sum of the other two terms. Hence the importance of carefully evaluating changes in r.

3. The model

3.1. Model atmosphere

We employ the single-column model of BR89. The model and its parameter sensitivity are thoroughly discussed in the original papers, so only a brief description is given here.

The model treats the Tropics as a region of horizontally uniform temperature and subsidence (upward motion is supposed to be concentrated within a population of mesoscale convective systems which at any time cover a negligible fraction of the total area). The upper and lower troposphere are separated by an inversion layer (the "trade inversion") of fixed depth but variable height. Below the trade inversion there is a convective boundary layer (CBL) subdivided into a cloud layer (containing a fixed fraction of non-precipitating clouds) atop a well-mixed subcloud layer, the boundary between the two being fixed by the lifting condensation level (LCL). The temperature and moisture structure of both clear and cloudy regions are determined using a mixing-line parameterisation (that is, values of conserved thermodynamic parameters at each level are given by a linear combination of the respective values at the surface and just above the inversion; the proportion of the mixture at each level is set by a single empirical constant). Above the trade inversion, the temperature follows a moist adiabat set by the subcloud equivalent potential temperature θ. The tropopause is at the level where this moist adiabat reaches 195 K; the temperature is fixed at this value at all higher levels.

Up to this point, we have specified the temperature profile throughout the atmosphere and the moisture profile in the CBL. To complete the model, we require a specification of the moisture profile above the CBL (which we refer to as the upper tropospheric humidity, UTH). In BR89, UTH is prescribed as follows. First, the inversion-top humidity, q_T, is determined as the saturation humidity at the level where the upper tropospheric moist adiabat reaches a specified temperature (-7°C). Given the inversion-top temperature, the saturation pressure departure $\mathscr{P}_T = p_T^* - p_T$ may be computed (p is pressure, p^* is saturation pressure, and subscript T indicates the level at the top of the inversion). The pressure departure at other levels is computed by linear interpolation between $\mathscr{P}_T$ and a prescribed tropopause value $\mathscr{P}_{TR}$. Finally, given the temperature profile, the moisture profile can be computed.

There is currently much debate around the issue of which mechanisms dominate the maintenance of the UTH profile, how they respond to changes in surface temperature and how best to parameterise them (Held and Soden (2000) give a recent overview). The parameterisation just outlined is not meant to reflect any particular mechanism, but simply to give a climate with more or less realistic values of UTH. Note however that it implies a fairly strong positive UTH feedback — that is, upper tropospheric specific humidity will increase quite sharply with temperature. As it turns out, UTH plays an important rôle in the problem under study here. To evidence this role, we will also conduct an experiment (see Section 4) in which this feedback is suppressed and UTH is kept constant as temperature changes.

The modeling assumptions outlined above allow the temperature and moisture structure of the entire tropical atmosphere to be determined by only three independent variables: the subcloud potential temperature θ_0 and specific humidity q_0 (which together define θ_e), and the inversion-top potential temperature θ_T. Equilibrium values of the three variables are determined by simultaneously solving the following set of equations:

$$F_s + F_l = \Delta N_{TR}, \tag{5}$$

$$\frac{c_p}{g}\,\omega_T(\theta_T - \theta_0) - \Delta N_T + F_s = 0, \tag{6}$$

$$\frac{L}{g}\,\omega_T(q_T - q_0) + F_l = 0, \tag{7}$$

$$F_l = \frac{L}{g}\,\omega_0(q_s - q_0), \tag{8}$$

$$F_s = \frac{c_p}{g}\,\omega_0(\theta_s - \theta_0), \tag{9}$$

$$F_s = \frac{\Delta N_B}{(1+k)}. \tag{10}$$

Eq. (5) is the moist static energy budget for the entire troposphere, while (6) and (7) are the static energy and moisture budgets for the CBL. Here ΔN_{TR} and ΔN_T are the radiative flux convergences integrated though the troposphere and the CBL respectively, ω_T is the inversion top subsidence rate, c_p is the specific heat of air and g is the gravitational acceleration. Eqs. (8) and (9) are bulk parameterisations for the surface latent and sensible heat fluxes respectively; q_s is the surface saturation specific humidity, computed using T_s and a surface pressure of 1012 hPa, and θ_s is the potential temperature at the sea surface. Finally, (10) represents the static energy budget for the subcloud layer. ΔN_B is the radiative flux convergence integrated through the subcloud layer. The parameter $k = 0.25$ accounts for entrainment from above; the equilibrium is not very sensitive to the value chosen. Following BR89, energy export to the extratropics is neglected in (5) and (6).

3.2. Surface wind parameterisation

To the above model, a parameterisation is added linking surface wind to atmospheric temperature, which allows ω_0 to be determined as part of the equilibrium. B99 proposes the following linear relation for the AM flux from the Tropics to the extratropics:

$$F_M = \overline{F_M} + a_M(\Delta Z_{500} - \overline{\Delta Z_{500}}). \tag{11}$$

Here $\Delta Z_{500} = Z_{500}^{trop} - Z_{500}^{extr}$ (where Z_{500}^{trop} is the 500 hPa geopotential height averaged between 0 and 30° latitude and Z_{500}^{extr} is the corresponding average between 30° and the pole) gives a gross measure of the meridional temperature contrast in the lower troposphere. F_M is the total AM transport across the 30° latitude line. Overlines indicate climatological values, and a_M is an empirical constant. In the present study, extratropical temperatures are held fixed by setting $Z_{500}^{extr} = \overline{Z_{500}^{trop}} - \overline{\Delta Z_{500}}$ so that (11) may be written

$$F_M = \overline{F_M} + a_M(Z_{500}^{trop} - \overline{Z_{500}^{trop}}). \tag{12}$$

Neglecting cross-equatorial transport, tropical momentum balance requires the sum of F_M and the surface momentum flux integrated between 0 and 30° to be zero. Using a bulk surface momentum flux parameterisation, we have

$$F_M = -\alpha\rho_a C_D|U|U = \frac{\alpha}{\rho_a C_D g^2}\,\omega_0^2, \tag{13}$$

where α is a geometric constant. Clearly, an increase in the mean meridional temperature gradient will result in increased surface winds.

4. Stability of the model SST

The aim now is to estimate the various terms in (3). This is done by computing a reference ocean–atmosphere equilibrium state and then studying the response of the atmosphere to SST perturbations.

4.1. Computation of reference and perturbed climates

A reference atmosphere–ocean equilibrium state for our model may be computed in the following steps:

- Firstly, we require that at equilibrium

$F_M = \overline{F_M}$; eq. (13) then sets the value of ω_0. The value of $\overline{F_M}$ is taken from observations.

- A value of T_s is now guessed. Eqs. (5)–(10) can then be solved to determine the corresponding atmospheric equilibrium. The equations are solved by first guessing values of θ_0, q_0 and θ_T, which allows detailed profiles of temperature and moisture to be computed as described in Section 3 (see BR89 for further details). The profiles are then fed into a radiative transport code (Chou aand Suarez, 1994, 1999) to compute radiative flux convergences ΔN_{TR} and ΔN_T. The equations can now be solved to obtain new values of θ_0, q_0 and θ_T. The process is repeated until the variables have converged to a given accuracy.
- At this point, the atmosphere is in equilibrium but the ocean, in general, is not: that is, eq. (1) will give a non-zero tendency for T_s. To determine its equilibrium value, T_s is allowed to evolve according to eq. (1), with the atmospheric equilibrium being recalculated at each time step. If our initial guess for T_s is near a stable equilibrium, the system will eventually converge to that equilibrium.

A reference equilibrium state was computed following this scheme and using the following parameter values: $\overline{F_M} = 3$ Hadley (roughly the observed value; 1 Hadley $= 1 \times 10^{18}$ kg m^2 s^{-2}), a solar zenith angle of 48.97° (chosen so as to give a surface temperature of exactly 300 K), and $F_o = -20$ W m^{-2} (roughly the observed ocean energy export to the extratropics); other parameters are as in BR89. In Table 1 we present equilibrium values of various quantities. The equations were iterated until the values of T_s, θ_0, q_0 and θ_T were accurate to 3 decimal places; this implies accuracy to 1 decimal place in the surface flux values.

The perturbed climate is computed by adding a small perturbation to the reference equilibrium T_s and recalculating the atmospheric equilibrium. To do this, ω_0 is first fixed and (5)–(10) are solved to get the new thermodynamic structure. A new value of ω_0 is then computed using (12) and (13); the procedure is iterated to convergence. The terms in (3) are estimated using the difference between perturbed and equilibrium surface fluxes. A +1 K perturbation is employed, but results are not sensitive to the size or sign of the perturbation. The response of the surface wind will of course depend on the value of a_M. For the northern hemisphere, B99 finds $a_M \approx 0.1$ Hadley m^{-1}; we explore the range 0–0.3 Hadley m^{-1}.

Table 1. *Values of key parameters in the reference equilibrium state. F_{SW} and F_{LW} are the net surface short- and longwave radiative fluxes (negative values imply upward flux)*

Parameter	Value	Unit
T_s	300.0	K
θ_s	299.0	K
θ_0	297.8	K
θ_T	306.5	K
q_s	22.47	g kg^{-1}
q_0	14.50	g kg^{-1}
q_T	4.55	g kg^{-1}
ω_0	0.0642	Pa s^{-1}
ω_T	0.0514	Pa s^{-1}
$\|U\|$	4.2	m s^{-1}
LCL	929	hPa
p_T	771	hPa
Z_{500}^{trop}	5862	m
F_{SW}	238.0	W m^{-2}
F_{LW}	−63.6	W m^{-2}
F_l	−130.3	W m^{-2}
F_s	−8.1	W m^{-2}
ΔN_B	−10.1	W m^{-2}
ΔN_T	−52.2	W m^{-2}
ΔN_{TR}	−138.4	W m^{-2}

The reference equilibrium was computed using the UTH parameterisation of BR89 (see Section 3). We examine the role of UTH feedback by considering two possible responses of UTH to SST perturbations. In the first case ("strong UTH feedback"), the perturbed atmospheric state is computed again using the BR89 parameterisation; in the second ("weak UTH feedback"), upper tropospheric specific humidity is kept fixed at its value in the reference equilibrium.

4.2. *Results*

4.2.1. Rôle of AM coupling. The three components of β (see eq. (3)) for the strong UTH feedback case are displayed in Fig. 1a as a function of a_M. To aid in the interpretation of these results, we plot in Fig. 2 the temperature, moisture and radiative profiles under equilibrium and perturbed conditions.

Let us focus first on the response when the AM coupling is switched off: that is, $a_M = 0$. As Fig. 1a shows, there are two positive contributions to β:

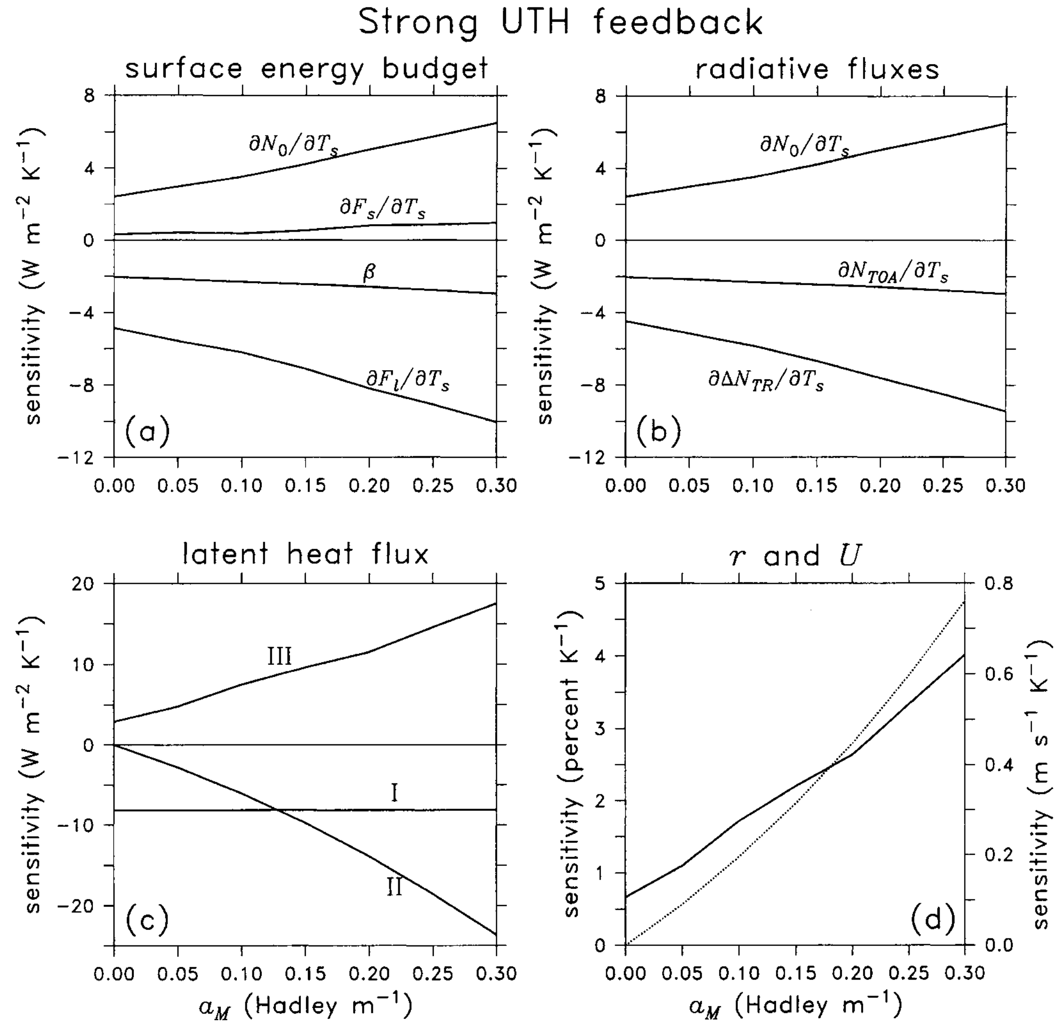

Fig. 1. The case with strong UTH feedback. (a) Sensitivity to T_s of the surface energy budget (see eq. (3)). (b) Sensitivity to T_s of the net surface and top-of-the-atmosphere radiative fluxes N_0 and N_{TOA}, and the net tropospheric radiative flux convergence ΔN_{TR}. (c) Sensitivity to T_s of the latent heat flux analysed into its three constituents (see eq. (4)): the Clausius–Clapeyron factor $(F_l/q_s)(\partial q_s/\partial T_s)$ (marked I), the wind factor $(F_l/\omega_0)(\partial\omega_0/\partial T_s)$ (marked II), and the relative humidity factor $(F_l/(r-1))(\partial r/\partial T_s)$ (marked III). (d) The sensitivity to T_s of subcloud relative humidity r (solid, left axis) and surface wind speed U (dotted, right axis).

a large one due to the radiative flux and a much smaller one due to the sensible heat flux. Both effects can be accounted for by the increase in subcloud humidity with SST (both absolute and relative, as Figs. 2d and 1d show). The destabilising response of the surface *longwave* flux is well known (Hartmann and Michelsen, 1993) and is due to the narrowing of the 8–12 μm water vapour window, which causes back (downwelling) radiation to increase more rapidly than the Stefan–Boltzmann law. This effect is somewhat offset by a negative feedback due to the shortwave flux, which decreases with increasing SST mainly because of higher absorption due to higher humidity throughout the troposphere (Fig. 2d). Increasing humidity also causes the LCL to drop, so that the subcloud layer becomes shallower; since the subcloud radiative flux gradients hardly

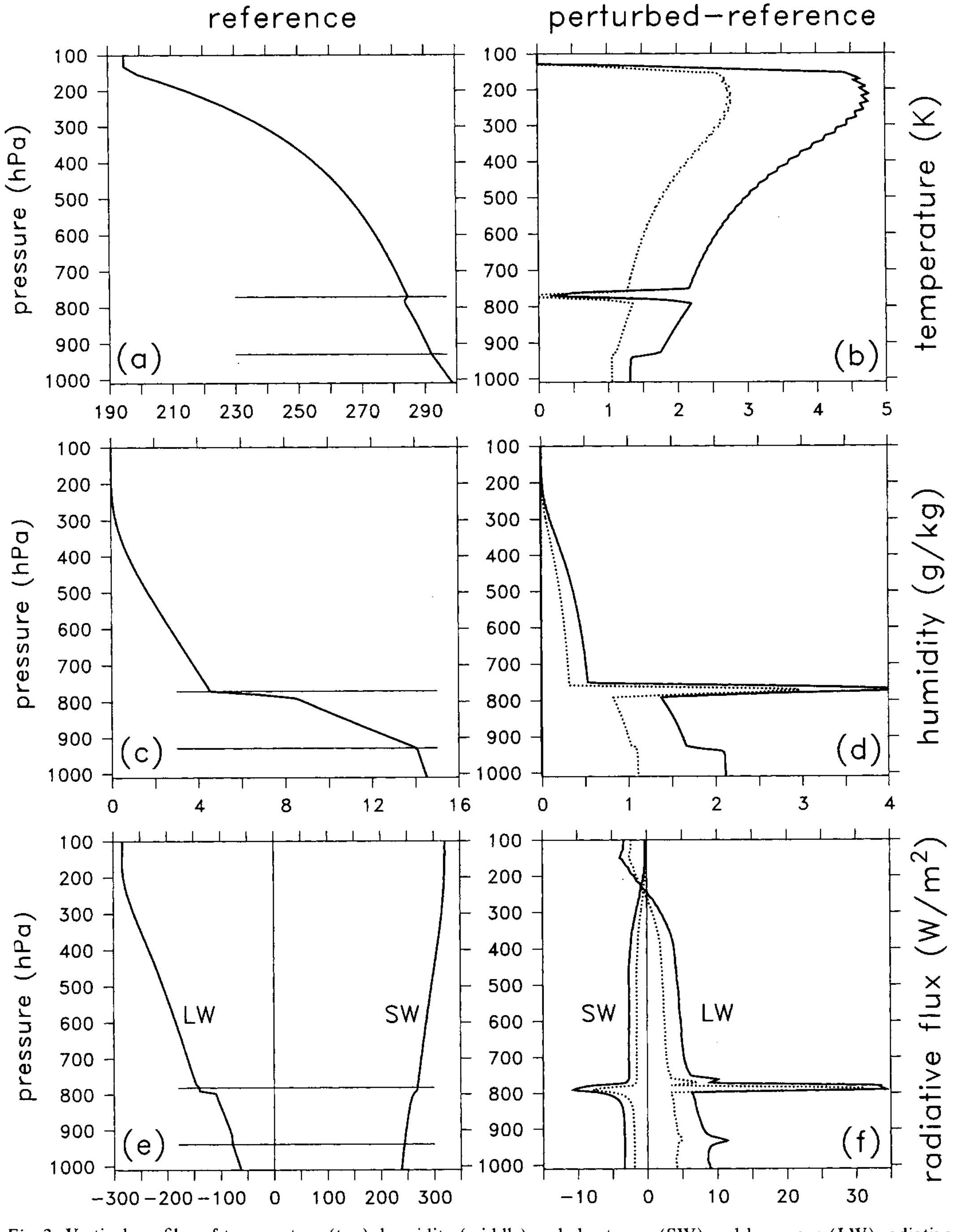

Fig. 2. Vertical profiles of temperature (top), humidity (middle) and shortwave (SW) and longwave (LW) radiative flux (bottom). The left-hand panels show profiles in the reference equilibrium state. The right-hand panels show the response of the profiles to a +1 K SST perturbation, in the case $a_M = 0$ (dotted) and $a_M = 0.3$ Hadley m^{-1} (solid). The thin horizontal lines in the left-hand panels indicate the extent of the cloud layer.

change (Fig. 2c,d), the net convergence ΔN_B decreases and hence, according to (10), so does F_s.

The SST is rendered stable by the evaporative cooling, which increases with SST providing a large negative contribution, so that overall β is negative. In Fig. 1c, the latent heat flux term is analysed into its three components (see eq. (4)). The figure shows that the relative humidity factor provides a destabilising contribution (indeed, Fig. 1d shows that r increases by about 0.75% for every 1 K increase in T_s). Physically, subcloud humidity increases both because of increased surface forcing and because ω_T decreases, so there is less entrainment of dry air from above the CBL. In Section 5, we will argue that in fact subcloud humidity *must* increase, because that is the only way to satisfy (5) and bring the atmosphere into equilibrium.

The requirement of tropospheric equilibrium implies a tight coupling between the surface and top-of-the-atmosphere (TOA) energy budgets. This can clearly be seen if we rewrite (5) in the form

$$F_s + F_l + N_0 = N_{TOA} \quad (14)$$

(N_{TOA} is the net radiative flux at the TOA), which simply states that any imbalance at the surface must be matched by a similar imbalance at the TOA. Comparing with (3), we see that

$$\beta = \frac{\partial N_{TOA}}{\partial T_s}.$$

Comparison of Figs. 1a and b shows that this is indeed the case. Further, examination of Fig. 2f shows that the response of N_{TOA} is dominated by the increase in outgoing longwave radiation (OLR). Thus we could just as well say that it is the OLR that renders the SST stable.

When AM coupling is switched on ($a_M > 0$), the magnitude of β clearly increases. Thus, AM coupling does indeed provide an "extra" stabilising effect. From the point of view of the surface budget, the extra stability is clearly due to the wind factor (marked "II" in Fig. 1c). From the point of view of the TOA budget, the extra stability can be attributed to a bigger response of the OLR (Fig. 2f). The latter occurs because, for a given value of SST, stronger surface wind results in an equilibrium with higher subcloud humidity; this entails higher θ_e and hence higher tropospheric temperatures, leading to higher OLR.

At $a_M = 0.1$ Hadley m^{-1}, the wind and Clausius–Clapeyron factors are about equal, in agreement with the results of the simple model of B99. However, as a_M increases, the relative humidity factor makes an increasingly destabilising contribution, so the extra stability gained by including AM coupling is not as high as it could be if r remained constant. This is the main difference with the results of B99.

4.2.2. Rôle of UTH feedback. Results for the weak UTH feedback case (in which UTH is held fixed as SST changes) are presented in Fig. 3. Looking at Fig. 3a, we see that the surface radiative and sensible heat flux terms are almost unchanged from their values in the strong UTH feedback case (Fig. 1a), while the latent heat flux term is much stronger. As a result, β has greater magnitude and the SST is much more stable. The latent heat flux is higher because the relative humidity factor plays a smaller role (Fig. 3c). Note indeed the smaller increase in r with SST (Fig. 3d). From another point of view, we can say that the increased stability of the system is due to the stronger response of N_{TOA} (Fig. 3b); this is due to the fact that, since UTH is fixed, the effective emission level remains unchanged, so that OLR increases faster with temperature (in the strong UTH feedback case, the increase in OLR due to increasing temperature is offset by the rise of the effective emission level, which lowers the effective emission temperature). The two points of view are of course complementary; we might summarise by saying that, in the weak UTH feedback case, the atmosphere manages to emit more radiation to space, so it can afford to absorb more energy from the surface in the form of latent heat. Note that in the strong (weak) UTH feedback case, β increases by about 1 (3) W m^{-2} K^{-1} over the range $a_M = 0$ to 0.3 Hadley m^{-1}: clearly, AM coupling has a greater stabilising effect when UTH feedback is weak.

5. A simplified model

The results above make clear the need to understand the dominant mechanisms which determine the equilibrium value of subcloud humidity. The mechanisms controlling q_0 (and hence r) are encoded in the equation set (5)–(10), but are not

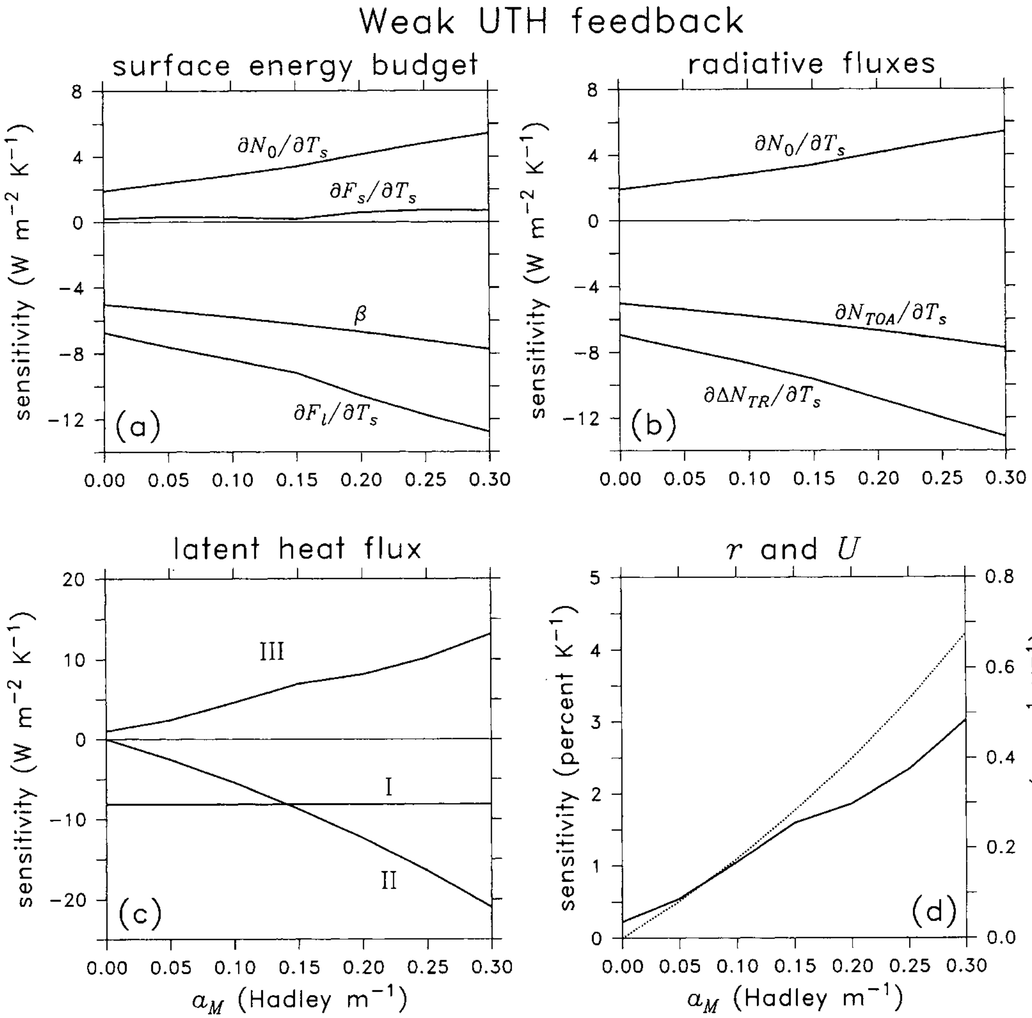

Fig. 3. As in Fig. 1 but for the case with constant specific humidity in the upper troposphere (weak UTH feedback).

immediately apparent because of the complicated interdependence of the three variables q_0, θ_0 and θ_T. It turns out that, by making two fairly robust approximations, the equations can be reduced to a significantly simpler form whose qualitative behaviour is more readily analysed.

The two approximations are as follows.

(i) The Bowen ratio $F_s/F_l = 0$. This is equivalent to requiring $\theta_0 = T_s$, so that the subcloud potential temperature exactly tracks the SST. The reason F_s is small compared to F_l is that it actually serves only to balance part of the small subcloud-layer radiative cooling. Climatologically, the subcloud layer is topped by a temperature inversion (Sarachik, 1978), so that relatively warmer air is entrained from above; i.e., the sensible heat flux at the top of the subcloud layer is downwards (in (10), this flux has been parameterised as $-kF_s$ with $k > 0$). Thus, sensible heat fluxes from the surface do not reach above the subcloud layer. The bulk of the net tropospheric radiative cooling is balanced by F_l.

(ii) ΔN_{TR} does not depend on θ_T. The role of θ_T is mainly to fix the depth of the CBL. Since OLR originates at levels higher than CBL top, while radiation exiting the bottom of the atmosphere originates at levels lower than CBL top, we do not expect ΔN_{TR} to be sensitive to θ_T.

Using the above two approximations, the set (5)–(10) can be rewritten as:

$$\frac{L}{g}\,\omega_0(q_s - q_0) = \Delta N_{TR}(T_s, q_0), \tag{15}$$

$$\frac{c_p}{g}\,\omega_0\,\frac{(q_s - q_0)}{(q_T - q_0)}(\theta_T - T_s) = -\Delta N_T(T_s, q_0, \theta_T), \tag{16}$$

having used (7) and (8) to eliminate ω_T in (6).

The crucial point is that the equilibrium value of q_0 is now given by the single equation (15). This equation states that q_0 is determined by requiring the latent heat flux to match the troposphere's net radiative cooling. Once q_0 is known, θ_T can be computed from (16).

Let us now consider what happens when, as in Section 4, we take the atmosphere–ocean equilibrium, perturb T_s, and allow the atmosphere to come back into equilibrium while keeping T_s fixed at its perturbed value. Clearly, before perturbing T_s, both (15) and (16) are satisfied, while after the perturbation this will no longer be the case. Bringing the atmosphere back to equilibrium means, in our context, adjusting the values of θ_0, q_0 and θ_T until (15) and (16) are again satisfied. However, we are assuming that q_0 is independent of θ_T, and that $\theta_0 = T_s$. Thus, we only really have freedom to play with q_0 in order to bring the atmosphere to equilibrium.

Let us see in detail how the two sides of eq. (15) evolve during the adjustment to equilibrium. We can conceptually subdivide the adjustment process into three steps, as shown in Fig. 4.

Step 1. Firstly, T_s is increased to its perturbed value while holding all atmospheric variables constant. F_l will increase sharply because of increasing q_s, while ΔN_{TR} will decrease somewhat due to greater upwelling longwave radiation at the surface.

Step 2. Next, θ_0 is increased to match the increase in SST, while q_0 and θ_T are held constant. This causes θ_e to increase, and as a consequence the entire troposphere warms. This means ΔN_{TR} (which is the net tropospheric cooling) must increase, unless the atmosphere is in a runaway greenhouse state. The extent to which it increases (i.e., the slope of the ΔN_{TR} curve) depends on the behaviour of the UTH. A strong UTH feedback — that is, a strong increase in UTH when surface temperature increases — will lead to a smaller slope, since the effective emission level will rise quickly so that OLR will increase only slowly with atmospheric temperature; this is indicated by the solid line. If the UTH feedback is weaker, then the slope will be steeper (dotted line). Meanwhile, F_l remains constant in the case $a_M = 0$ (solid line), since neither q_0 nor the surface wind are affected. If $a_M > 0$ (dotted line), surface wind will increase, and hence also F_l.

Step 3. Now we adjust q_0 until eq. (15) is satisfied. The *only* way to do this is to *increase* q_0. This makes F_l drop (because $q_s - q_0$ decreases), while ΔN_{TR} will rise since higher q_0 means higher θ_e and hence higher temperature throughout the troposphere. The final atmospheric equilibrium will therefore always have a more humid subcloud layer than the reference equilibrium. Thus, the humidity factor will always have a destabilising influence (that is, it will always act to reduce the evaporative cooling).

As an epilogue, θ_T is adjusted according to (16).

We emphasise two aspects of this analysis. Firstly, we stress again that ΔN_{TR} can only increase with θ_e. In other words, the net radiative cooling must increase if the mean tropospheric temperature increases, otherwise we would find ourselves in an unstable "runaway greenhouse" situation. If ΔN_{TR} were to decrease, then Fig 4 makes it clear that F_l would also decrease (though T_s has increased), so that the ocean temperature would also run away to high values. The point is that we cannot consider the stability of the ocean as a separate issue from that of the atmosphere.

Secondly, it is clear from Fig. 4 that the greater the value of a_M, the larger the final value of F_l: thus, AM coupling will always provide some "extra" stability (as compared with the $a_M = 0$ case). However, the size of this extra stability depends not only on a_M but also on the strength and sign of the upper tropospheric water vapour feedback, which determines the slope of the ΔN_{TR} curve. The more UTH increases with SST, the weaker the stabilising effect of AM coupling.

6. Conclusions and discussion

The role of AM coupling (that is, the positive correlation between surface wind and the mean

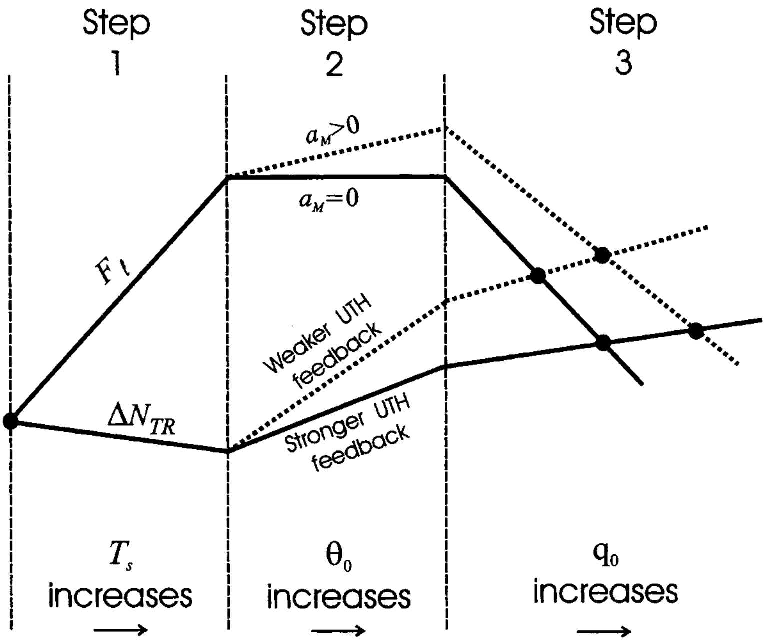

Fig. 4. Schematic depiction of the atmospheric adjustment to equilibrium under a positive SST perturbation. Black dots indicate equilibrium states; the leftmost dot is the reference equilibrium. See text for details.

meridional temperature gradient) in stabilising the tropical SST has been studied using a model in which subcloud humidity is internally determined rather than being arbitrarily specified. The only constraint on the model is that the atmosphere be in equilibrium at all times. Results show that AM coupling indeed provides a stabilising feedback, but also leads to an increase in subcloud humidity which tends to diminish the strength of this feedback. We have also formulated a simplified version of the model which permits a detailed analysis of the processes determining the value of subcloud humidity. The analysis leads us to two main conclusions.

(i) AM coupling will always enhance the evaporation rate and hence provide a stabilising feedback on SST under tropical conditions;

(ii) the strength of this feedback depends in an essential way on the strength and sign of the upper tropospheric water vapour feedback. The current uncertainty concerning the latter feedback entails a corresponding uncertainty in the former.

We emphasise that these conclusions strictly concern only the stability of the tropical SST with respect to tropical, perturbations such as might arise, say, from changes in tropical cloud cover. It bears pointing out, however, that AM coupling could have a destabilising effect on the extratropics. Consider for instance a cooling perturbation concentrated in the extratropics. It will tend to increase the mean meridional temperature gradient and hence strengthen surface winds. If the extratropics were to respond to increased surface wind by cooling (as the Tropics do), then we would have a positive feedback effect. Clearly, we cannot answer such questions without an appropriate model for the extratropical climate. One thing we can say, however, is that we expect an extratropical cooling (such as that due to increased albedo during glacial times) to increase surface winds in the Tropics and hence to have a cooling effect there. Indeed, in a coupled GCM study of the Last Glacial Maximum, Bush and Philander (1998) found that the chief mechanism for Tropical cooling was increased surface wind.

The atmosphere in the present model is assumed to be everywhere subsiding; in effect, we are assuming that the compensating ascent occurs in a region of negligible area. A number of investigators (Pierrehumbert, 1995; Miller, 1997; Larson et al., 1999; Nilsson and Emanuel, 1999; Clement and Seager, 1999; Seager et al., 2000) have studied the more general case where the rising branch of the circulation occupies some finite area. This introduces a new external parameter (the area ratio of rising and subsiding branches), and it would be interesting to examine the behaviour of our results as a function of this parameter. The papers of Larson et al. (1999) and Seager et al. (2000) both show that SST decreases with increasing surface wind, just as in BR89. This essentially means that increased surface wind implies increased OLR, which in view of the discussion in Sections 4 and 5 suggests that AM coupling will also play a stabilising role in those models.

The models of Larson et al. (1999) and Nilsson and Emanuel (1999) are also of interest because UTH is explicitly computed. Though it is difficult to ascertain the realism of these parameterisations, given the current lack of empirical and theoretical knowledge regarding upper tropospheric moistening, it would be of interest to see if there are situations in which increasing surface wind can increase UTH: in that case, AM coupling could have a destabilising effect in the Tropics.

At an early stage (eq. (3)), feedbacks due to changes in ocean heat export were disregarded, as these are unrelated to the issue of evaporation. However, AM coupling is likely to play a significant role here, since increasing surface stress will increase both the gyre circulations and the Ekman transports (see for instance Cessi, 2000).

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