Thermostats, Radiator Fins, and the Local Runaway Greenhouse*

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ABSTRACT

The author has reconsidered the question of the regulation of tropical sea surface temperature. This has been done in general terms through consideration of the tropical heat budget and in specific terms through consideration of an idealized radiative–dynamic model of the tropical general circulation. It is argued that evaporation on its own cannot provide an effective regulating mechanism. Clouds cannot serve as regulators unless there are substantial departures from the observed cancellation between cloud greenhouse and cloud albedo effects. In particular, it is shown that the prediction by Ramanathan and Collins of highly stable tropical climates is based on an inconsistent set of assumptions about the behavior of the atmospheric heat transports. When the heat transports are treated in a consistent manner, clouds are found to have little impact, and the tropical climate can be quite sensitive to radiative perturbations.

It is found that the main determinant of tropical climate is the clear-sky water vapor greenhouse effect averaged over the entire Tropics. In the absence of dry "radiator fins" maintained by subsidence, the tropical temperature would tend to fall into a runaway greenhouse state that could be stabilized only by heat export to the extratropics. Some speculative results on sensitivity of the climate to perturbations are presented. Determination of the relative area of dry and subsiding versus moist and convecting regions of the Tropics, and of the degree of dryness of the subsiding regions, are identified as key unsolved problems concerning the tropical climate.

1. Introduction

Determining the temperature of the warm tropical oceans stands as one of the central problems in the study of climate. The Tropics account for a substantial fraction of the earth's surface, providing consistently equable environments that have been host to biologically diverse ecosystems during much of the earth's history. Moreover, in the annual average the Tropics are the locus of greatest incident solar radiation; the efficiency with which they retain this heat determines the pool of energy available for export to higher latitudes. The fluctuations of climate through time can be thought of as resulting from an uneasy competition between the tropical solar furnace and the ponderous thermal flywheel provided by high-latitude ice sheets. ·

There is evidence that the maximum tropical sea surface temperature (SST) was never at any time during the past 100 million years much greater than it is at present (Crowley and North 1991), despite

substantial fluctuations in continental position, atmospheric CO_2, and extratropical climate. During the Cretaceous, though the extratropics were considerably warmer than at present, the Tropics may not have been any warmer than they are currently (Sellwood et al. 1994). On the cold side, the indication from foraminifera assemblages (CLIMAP 1976) has generally been taken to mean that during the last glacial maximum the tropical oceans were no more than 1 K cooler than at present. This finding is problematic in light of tropical mountain snow-line data (Rind and Peteet 1985; Rind 1990) suggesting coolings of around 4 K in the midtroposphere, which could translate into a surface cooling of perhaps 2 K using conventional models (Betts and Ridgway 1992), though speculative models of the tropical lapse rate (Sun and Lindzen 1993a) allow weaker cooling. Recent data from coral cores (Guilderson et al. 1994), however, indicate a 3–4 K cooling of the ocean surface. Fluctuations about the present climate provide some evidence for stability, in that the atmosphere–ocean system accomplishes a major reorganization of convection and oceanic heat fluxes in the course of El Niño without ever increasing the maximum Pacific SST much beyond 302 K (Philander 1990). From an observational standpoint, the question of the resilience of the tropical climate is rather unsettled, particularly on the cold side. If it could be stated with confidence that the tropical climate was insensitive to perturbations, the whole enterprise of modeling

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climate change—notably estimation of the amount warming expected from doubling CO_2—would become much simpler. The Tropics could then be taken as an essentially immutable boundary condition for the extratropical climate.

Comprehensive theoretical treatments of the factors governing tropical SST have been given by Sarachik (1978), Betts and Ridgway (1989, 1992), and Sun and Lindzen (1993), among others. Sensitivity of the tropical climate has also been extensively studied using full general circulation models (e.g., Broccoli and Manabe 1987). On the opposite extreme, there has been a recurring desire to isolate very simple, dominant aspects of the physics that can serve in themselves as stabilizing feedbacks. Early attention settled on evaporation as a stabilizing effect (Newell 1979), owing to the strong temperature dependence implied by the Clausius–Clapeyron relation. More recently, Ramanathan and Collins (1991, hereafter RC) have used satellite radiation budget data to argue in favor of cloud albedo effects as the dominant stabilizing influence, a proposal they refer to as the "thermostat hypothesis." There has been much discussion surrounding these two proposals (cf. the review in Waliser and Graham 1993), but in fact it can be shown on quite elementary grounds that neither one is viable. Explaining why this is so is one of the aims of the present work. Much of the discussion of the thermostat hypothesis has focused on the question of whether cloud optical properties are controlled by local SST as assumed by RC. Information germane to this issue can be found in Fu et al. (1992), Hartmann and Michelsen (1993), and Waliser and Graham (1993), but the arguments put forth in the present work are entirely separate from this aspect of the debate. In the present work, we probe the effect of clouds on tropical climate conditional on the assumption that cloud albedo and cloud greenhouse effects cancel in the top-of-atmosphere radiation budget, and we conclude that clouds are not the key players in determining surface temperature. This contrasts with the conclusions of RC, who argue that clouds would have a strongly stabilizing effect even in the presence of an exact cancellation.

Although we throw some cold water on the notion that clouds or evaporation regulate the tropical climate, we do not offer an unambiguous stabilization mechanism in its place. Instead, by examining the behavior of a simple model embodying most of the "zero-order" thinking about the basic operation of the Tropics, we endeavor to identify the effects that are important enough to be considered as candidates for stabilizing or destabilizing feedbacks. If nothing else, this enterprise helps to force into the open the various assumptions that must be made in estimating climate sensitivity using observations or general circulation models. Attention will be restricted to situations in which the atmosphere–ocean system can be regarded as being in equilibrium. Thus, our results apply to sufficiently long time averages of the tropical SST. Given the rather

small seasonal and interannual SST fluctuations in the Tropics, this is unlikely to be a fatal limitation. In any event, a thorough understanding of the climatological average is a necessary starting point for any treatment of the fluctuations about the average.

There is very little in this paper that is completely new, though we may perhaps claim some novelty in the way we have put together the big picture. Many of the matters we discuss have appeared in one form or another in the literature, but even if some of these are "well known," judging from the nature of the continuing controversy surrounding tropical SST regulation their significance is not universally appreciated. A number of basic concepts in this category are established in section 2. In section 3, the energy budget of the Tropics as a whole is used together with some plausible simplifying assumptions in order to estimate lateral atmospheric heat fluxes. This estimate serves as the basis for a precise treatment of the factors governing climate sensitivity. In section 4, the thermostat hypothesis of RC is reevaluated in light of what has been learned in the preceding two sections. In section 5, the consequences of the general ideas developed in section 3 are explored quantitatively within a minimal two-box model of the tropical general circulation. For the convenience of the reader, we have provided an appendix containing a glossary of the symbols used in this paper.

The picture emerging from our work is that the disposition of dry air pools, and the degree of their dryness, is the prime internal determinant of the stability of the tropical climate. However, it will also be seen that heat transport out of the Tropics has a significant cooling effect.

2. Preliminary considerations concerning the basic operation of the tropical climate

In this section we will review the basic physics of the runaway greenhouse, explore the prospects for stabilizing it by cloud feedbacks, and examine the joint role of clouds and evaporation in determining sea surface temperature *relative to the overlying air temperature*. The starting point is a detailed consideration of the energy budget of a column of the atmosphere–ocean system. The terms entering into the budget are summarized in Fig. 1. The following notations are introduced for the various heat fluxes:

$F_v(\infty)$ Net vertical flux into the top of atmosphere

$F_v(0)$ Net vertical flux out of the bottom of atmosphere

F_{ah} Net horizontal atmospheric heat transport (latent + sensible) into the column (per unit cross sectional area)

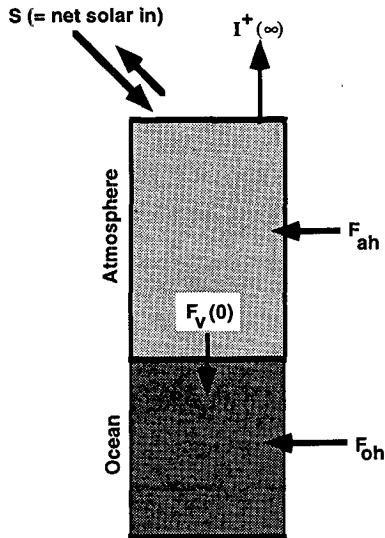F_{oh} Net horizontal oceanic heat transport into the column (per unit cross sectional area).

FIG. 1. The energy budget of a column
of the atmosphere–ocean system.

Each of these has the dimensions of energy flux[1] (e.g., W m^{-2}), and each is a function of latitude, longitude, and time; $F_v(0)$ is also the net flux into the ocean surface. The rate H of energy accumulation by the column is then

$$H = F_v(\infty) + F_{ah} + F_{oh}. \qquad (2.1)$$

Satellite observations, such as the Earth Radiation Budget Experiment (ERBE), measure $F_v(\infty)$, which is purely radiative. Let $I^+(\infty)$ be the upwelling infrared flux measured at the top of the atmosphere (commonly called "OLR") and S be the absorbed solar radiation similarly measured. The ERBE analysis (Ramanathan et al. 1989) subdivides these further into contributions from clear and cloudy skies as follows:

$$S = S_c + C_s, \qquad (2.2a)$$

$$I^+(\infty) = I_c^+(\infty) - C_l, \qquad (2.2b)$$

where C_s and C_l are the cloud shortwave and longwave forcings; S_c is an estimate of what the absorbed solar radiation would be in the absence of clouds, obtained by extrapolating from clear-sky pixels; and similarly, $I_c^+(\infty)$ is an estimate of what the OLR would be in the absence of clouds—it does not represent the emission from the clear-sky pixels of the scene alone but rather the longwave flux the scene would have if *all* pixels radiated at the same rate as the clear-sky pixels.

The top-of-atmosphere flux can thus be written

$$F_v(\infty) = S_c - I_c^+(\infty) + (C_s + C_l). \qquad (2.3)$$

Empirically, C_s and C_l sum to nearly zero in the present-day tropical climate (Ramanathan et al. 1989; RC). This remarkable cancellation holds not only for long-term averages but also for climate fluctuations on time-scales as short as a month, if not shorter. The physical basis of the cancellation is so far unexplained, and the circumstances under which the cancellation will continue to hold in perturbed climates are unknown. Though we will carry along the $(C_s + C_l)$ term in the budgets so far as practical, our discussion of cloud feedbacks in the present work will be predicated on the assumption that $(C_s + C_l) = 0$. We thus address the question of whether clouds can stabilize the tropical climate even if the top-of-atmosphere cancellation continues to hold. Stabilization mechanisms in which C_s becomes dominant in the perturbed climate are of course possible, but to posit a kind of cloud feedback that is utterly outside the range of behavior experienced in the course of natural variability requires a correspondingly higher degree c confidence in the cloud prediction scheme.

Suppose for the moment that $F_{ah} = F_{oh} = 0$. Then, the state of the climate is determined by balancing absorbed solar radiation with OLR. In Fig. 2, we show the OLR computed for the family of temperature profiles:

$$T(\zeta) = T(0) - \gamma\zeta \quad \text{for} \quad \zeta < \zeta_T$$

$$= T(0) - \gamma\zeta_T \quad \text{for} \quad \zeta > \zeta_T, \qquad (2.4)$$

where $\zeta = -h \ln(p/p_{\text{surf}})$ and p is the pressure. With $h = 7$ km, $\gamma = 6.2$ K km^{-1} and $\zeta_T = 16$ km, the profile (2.4) gives a reasonably good fit to the present-day tropical profile. Relative humidity at all levels was held
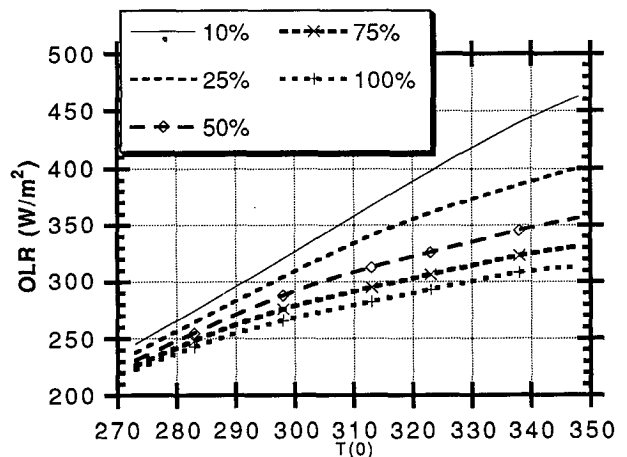

FIG. 2. Clear-sky OLR computed for various relative humidities, using the NCAR CCM radiation code and present-day CO_2. Results are shown as a function of low-level air temperature $T(0)$. See text for details concerning the temperature profile.

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[1] The terms we refer to as "horizontal transports," like F_{ah}, are, strictly speaking, the vertical integrals of the divergence of the horizontal heat fluxes. For the sake of economy of expression, we will refer to these loosely as "fluxes" or "transports."

fixed at the indicated values as $T(0)$ was varied, the CO_2 concentration was set to its present-day value, and cloud-free conditions were assumed. For the purposes of the radiation calculation, surface temperature was specified as $T(0) + 1$ K, but owing to the optical thickness of the moist atmosphere this has little effect on the results. These results were computed with the NCAR-CCM2 radiation code (Kiehl and Brieglieb 1992). Note that with appreciable relative humidity, the OLR increases very little with increasing temperatures and appears to level off at large temperatures. As $T(0)$ increases, the increased moisture content of the atmosphere shifts the radiating level to higher altitudes, which keeps the OLR from increasing much. This is the essence of the runaway greenhouse effect that has been extensively studied in connection with planetary evolution problems (Kasting 1967; Ingersoll 1969; Kombayashi 1967; Nakajima et al. 1992; Renno et al. 1994). We have exhibited it in a rather crude form, owing to the employment of a fixed temperature profile, but when the profile is consistently computed using a radiative–convective model, as in some of the studies cited, an absolute upper bound on the OLR that can escape from a moist atmosphere is nonetheless found. The alterations of lapse rate with temperature, of course, affect the shape of the OLR curve and have a significant affect on the surface temperature associated with a given OLR (e.g., Betts and Ridgway 1992; Sun and Lindzen 1993). The lapse rate feedback does not, however, alter the conclusion of a threshold of around 320 W m^{-2} for models that maintain moderate to high relative humidity aloft.

From Fig. 2 we see that if the full annual-mean equatorial insolation of 420 W m^{-2} were absorbed, $T(0)$ would run away to temperatures in excess of 340 K for any relative humidity greater than 25%. Even if S_c is reduced to 370 W m^{-2} to account for the mean clear-sky albedo in the Tropics (estimated from Fig. 2 in Stephens and Greenwald 1991), the temperature would run away for relative humidities as low as 50%. Considered locally, the present-day Tropics would thus be in a runaway state (or nearly so) so long as it is sufficiently close to saturation. Clouds do not alter this conclusion, because insofar as $C_s + C_l \approx 0$ in the Tropics, the reduction in solar absorption is compensated by an equal reduction in OLR. In order to stabilize the tropical run away, one must appeal mainly to lateral heat transports out of the moist regions. Satellite observations show clear-sky OLR of 300 W m^{-2} or less over the warmest tropical oceans, confirming the inability of the warmest oceans to get rid of the absorbed solar radiation locally (Raval and Ramanathan 1989; Stephens and Greenwald 1991).

Because of the optical thickness of the moist tropical atmosphere, the OLR is insensitive to the underlying surface temperature. Hence, the problem of determining the atmospheric temperature decouples from the surface temperature problem. The low-level air temperature $T(0)$ can be obtained first from the preceding considerations, once the lateral atmospheric heat transport out of the column is specified. Calculation of surface temperature in terms of $T(0)$ completes the problem.

Assuming the surface to be in thermal equilibrium, its temperature is determined by the surface energy budget

$$F_v(0) + F_{oh} = 0, \qquad (2.5a)$$

where

$$F_v(0) = \alpha S_c + C_s^* + [I^-(0) - \sigma T_s^4] - E - F_{sens}, \qquad (2.5b)$$

and α is a shortwave absorption coefficient representing the portion of the clear-sky top-of-atmosphere solar flux that reaches the ground; recall that albedo is already taken into account in the definition of S_c. The effect of clouds on the *surface* insolation is represented by C_s^*. It could in principle differ from C_s, owing to the effects of solar absorption within the cloud or in the subcloud air. The downwelling infrared flux impinging on the surface is $I^-(0)$, T_s is the surface temperature, E is the evaporative heat flux, and F_{sens} is the sensible heat flux.

To see how the balance operates, it is necessary to fill in some details concerning the behavior of the various terms in (2.5b). Evaporation is given by a bulk aerodynamic formula (Peixoto and Oort 1992)

$$E = \rho L C_d u^* q^*, \qquad (2.6a)$$

where ρ is the air density in the boundary layer, L is the latent heat of vaporization, C_d is the drag coefficient, u^* is the characteristic velocity fluctuation, and q^* is the characteristic scale of fluctuation of the water vapor mass mixing ratio. The scale q^* can be estimated using

$$q^* = q_{sat}(T_s) - r q_{sat}[T(0)], \qquad (2.6b)$$

in which q_{sat} is the saturation mixing ratio and r is the typical relative humidity of air entrained into the boundary layer. For the sake of the present argument concerning the surface budget, we will accept the assumption that cloud shortwave forcing is a function of surface temperature, writing

$$C_s^* = -a(T_s - 300 \text{ K}) \quad \text{for} \quad T_s > 300 \text{ K}. \qquad (2.7)$$

We use $a = 34$ W m^{-2} K^{-1}, based on recent direct observations of the surface insolation in the central equatorial Pacific (W. Collins 1993, personal communication). Further, C_s^* is set to zero below the 300 K convective threshold and saturates to $-\alpha S_c$ at the temperature where all the solar radiation is blocked. To simplify the discussion, we set F_{sens} to zero.

The net radiation heating the surface is $\{\alpha S_c + C_s^* + [I^-(0) - \sigma T_s^4]\}$, and the sum of this with F_{oh} must balance the evaporation if the surface is to be in equi-

librium. In Fig. 3, we show the radiation plus ocean aggregate with and without clouds, and the evaporation, as functions of T_s. The value $I^-(0) = 393$ W m^{-2} was computed from the NCAR radiation code with T prescribed according to (2.4), $T(0)$ held fixed at 300 K, and a relative humidity of 75% throughout the atmosphere; it is quite insensitive to the temperature and humidity above the lowest kilometer. The other parameters are the following: $S_c = 370$ W m^{-2} (to allow for clear-sky albedo), $\alpha = 0.8$, $F_{oh} = -30$ W m^{-2}, $u^* = 4$ m s^{-1}, $r = 80\%$, and $C_d = 0.002$. With no evaporation or cloud effects, the surface would equilibrate at the torrid temperature of 328 K. If clouds were allowed but evaporation was somehow suppressed, the temperature would settle down at 304.8 K, where the clouds almost completely shut off the solar radiation, assuming extrapolation of (2.7) to such extreme conditions to be valid. However, evaporation alone would limit the temperature to 302.8 K, and the joint effects of evaporation and clouds would only reduce this figure to 301.5 K. While clouds indeed accomplish some reduction in surface temperature *for fixed T(0)*, their impact is not overwhelming. Of course, the numbers are somewhat sensitive to the evaporation model employed. Increasing u^* or reducing r, for example, reduces $T_s - T(0)$, and further decreases the impact of clouds on T_s, but at the expense of increasing evaporation—perhaps beyond the range that is consistent with observations. At this point, however, it is not our goal to seek the most precise agreement between the model and observations but rather to illustrate fundamental principles, which are firmly rooted in the physics of evaporation.

Essentially, evaporation acts as a buffer and takes up the slack between the net radiation received by the surface and the heat flux needed to keep the surface temperature near the low-level air temperature. The principal effect of introducing clouds is to reduce the evaporation from 182 to 135 W m^{-2}. Evaporation would have a similar buffering effect against changes in F_{oh}. The buffering effect of evaporation is also evident in the results of Stephens and Webster (1984) and Lindzen et al. (1982), who employed more realistic radiative–convective and boundary-layer physics. For that matter, the argument presented in Fig. 3 is the same as that put forth by Newell (1979), apart from the introduction of clouds. Newell (1979) was essentially correct, except that the argument constitutes a theory of the temperature differential $T_s - T(0)$, rather than a self-contained theory of sea surface temperature itself.

The tight coupling between surface temperature and air temperature over warm oceans is well supported by data. In Fig. 4, we show the monthly-mean difference between surface temperature and air temperature 2 m above the surface for March 1989. The data was taken from European Centre for Medium-Range Weather Forecasts (ECMWF) analyses. The largest differences, in excess of 10 K, are found over the daytime Sahara;

this is indicative of what would prevail if evaporation were somehow shut off, leaving sensible heat flux to do all the work. Fairly large differences are also encountered over the Gulf Stream and Kuroshio, which show that evaporation cannot entirely keep up with the extremely large oceanic heat fluxes in these regions. Over most of the tropical ocean, however, the temperature difference is 1 K, or less. This is true despite the substantial variation in cloud albedo over the Tropics, indicating that there is no tendency for the surface temperature excess to run away in the absence of clouds. There is a small region in the central subtropical Pacific (marked with an arrow in Fig. 4) 'where the temperature difference reaches 3 K. This is at the edge of the Tropics, where horizontal air temperature gradients and seasonal air temperature variations are very strong and may simply represent a transient effect due to the long relaxation time of SST.

In certain circumstances, the tropical boundary layer can be capped by a stable inversion, referred to as the "trade inversion." Formation of a trade inversion is particularly favored in regions of undisturbed subsidence (Emanuel 1991; Betts and Ridgway 1989 and references therein). In this circumstance, the preceding argument for tight coupling of SST and overlying air temperature still applies, but the boundary layer itself becomes somewhat decoupled from the physics determining the temperature profile over the rest of the troposphere. In consequence, the trade inversion, when it exists, allows the lowest-level air temperature to be a few degrees cooler than might be expected by an extrapolation from aloft using (2.4) or some other prescription (e.g., the moist adiabat). This phenomenon will resurface in section 5, in connection with our discussion of the cold pool SST.

3. Horizontal atmospheric heat transports and their effect on climate stability

We now pick up the thread where we left it at (2.3) and bring lateral heat transports back into the picture. The essence of the argument in this section is quite simple. First, equilibrium of the surface energy budget implies that solar energy delivered to the ocean surface, minus the part that is carried away by oceanic transports, is returned to the bottom of the atmosphere by whatever combination of fluxes serves to keep the budget balanced. Second, dynamical transports in the atmosphere prevent appreciable horizontal variations of the tropical atmospheric temperature, outside of a thin boundary layer. In consequence, the atmospheric temperature can be determined by satisfying the atmospheric column energy budget averaged over the entire Tropics, without reference to geographical variation of the terms in the budget. These two assumptions almost immediately lead to (3.9), which gives the sensitivity of tropical temperature to changes in radiative or oceanic forcing. We have elaborated the argument,
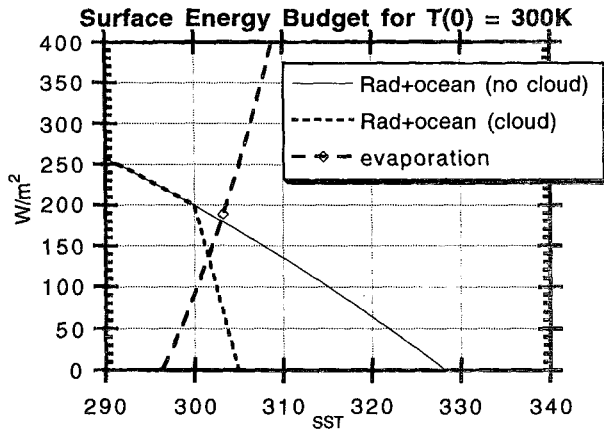
Surface Energy Budget for T(0) = 300K

FIG. 3. The surface energy balance for $T(0) = 300$ K, with and without clouds. See text for specification of the parameters of the calculation.

however, to allow for a discussion of the nature of the lateral fluxes that enforce near homogeneity of the atmospheric temperature. The assumptions concerning the temperature profile and shape of the OLR curve, introduced in the discussion following (2.3), play no role in the present section, though we shall have occasion to make use of them again in section 5.

The key to estimation of the atmospheric transport F_{ah} is a careful evaluation of the atmospheric heating.

Let Q_v be the energy added to the atmospheric column due to the convergence of the vertical flux. Thus,

$$Q_v = F_v(\infty) - F_v(0). \qquad (3.1)$$

Note that this includes radiative heating (visible and longwave) and latent heating due to local evaporation, plus latent heat storage due to unprecipitated evaporation from the local ocean surface. It does not include latent heat from imported water vapor or dynamical heating due to adiabatic compression of the column; these are accounted for in F_{ah}. Using (3.1) the column heat budget can be rewritten

$$H = Q_v + F_{ah} + [F_v(0) + F_{oh}]. \qquad (3.2)$$

Now we shall begin to introduce some simplifying assumptions. First, we assume that the system and all of its individual components are in equilibrium. This assumption is appropriate to study of the long-term climate of the Tropics, and the fluxes need only be reinterpreted as time averages over a suitably long interval. Because the atmosphere is in equilibrium, there is no transient storage of unprecipitated water. The evaporation into the column must either precipitate out (in which case it shows up as latent heat release) or be carried away (in which case it is accounted for in F_{ah}). Equilibrium of the ocean implies, in fact is defined by, $[F_v(0) + F_{oh}] = 0$.

In equilibrium, the column energy budget becomes
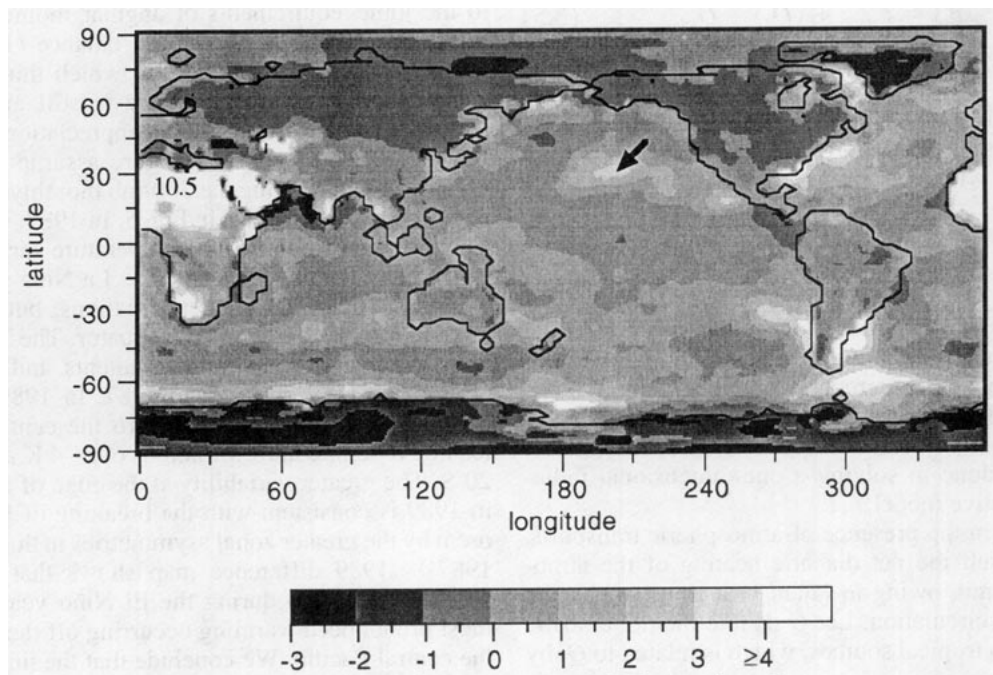
March 1989 (surf. temp)-T(2 m. above)

Data at 12Z

FIG. 4. March 1989 difference between 2-m air temperature and surface temperature, from ECMWF analyses.

$$H = Q_v + F_{ah}, \qquad (3.3)$$

which is naturally identical to the energy budget for the atmosphere alone. Further, the atmospheric heating due to local sources becomes

$$Q_v = S_c - I_c^+(\infty) + (C_s + C_l) + F_{oh}, \qquad (3.4)$$

where Q_v is the heating that forces the atmospheric circulation. If $Q_v = 0$ everywhere, then the climate consists of local radiative–convective equilibrium in each column, with no atmospheric circulation. *To the extent that $(C_s + C_l)$ is small throughout the Tropics, clouds do not affect the column-integrated forcing of the atmosphere.* Clouds could indirectly affect the atmospheric circulation by changing the vertical distribution of the heating, but this is a more subtle effect whose investigation would require a more detailed treatment of atmospheric dynamics than we wish to undertake in the present work.

Next, we introduce the assumption of *infinitely efficient atmospheric heat transport.* By this we mean that the heat transport takes on whatever value necessary to keep the atmospheric temperature horizontally uniform, no matter what differential forcing Q_v nature should see fit to produce. If we think of the tropical atmosphere as a thermally conducting slab, the efficient transport assumption amounts to the limit of infinite horizontal conductivity. The efficient heat transport assumption allows all the details of atmospheric dynamics to be avoided and replaced by the formula

$$F_{ah} = F_{aexp} + \langle Q_v \rangle - Q_v, \qquad (3.5)$$

where angle brackets represent an average over the Tropics and F_{aexp} is a constant independent of latitude and longitude. On averaging (3.5), it is seen to be the average energy flux exported from the Tropics by the atmosphere, with negative values signifying cooling of the Tropics. Expression (3.5) has the effect of replacing (3.3) by

$$H = \langle Q_v \rangle + F_{aexp}. \qquad (3.6)$$

Because this expression is independent of latitude and longitude it allows a solution with horizontally homogeneous atmospheric temperature. The solution is obtained by adjusting a single temperature profile $T(\zeta)$ so as to zero out the single quantity given by (3.6), much as is done in solving a one-dimensional radiative–convective model.

Note that in the presence of atmospheric transports, Q_v is not itself the net diabatic heating of the atmospheric column, owing to latent heat transport by the atmospheric circulation. Let Q denote the net diabatic heating from tropical sources, which is related to Q_v by the expression

$$Q = Q_v + (P - E), \qquad (3.7a)$$

where P is the latent heat release due to precipitation in the column and E is the evaporative heat flux into the column from the ocean surface. The atmospheric latent heat transport is $(P - E)$, whence we can write

$$F_{ah} = (P - E) + F_{ahs}, \qquad (3.7b)$$

where F_{ahs} is the atmospheric sensible heat transport. On substituting (3.7a) and (3.7b) into (3.5), it is seen that the efficient transport "ansatz" can be equally well obtained by positing that the *sensible* heat transport redistributes Q. In this sense, it is immaterial whether the imported latent heating is considered a forcing or a response, so long as the atmospheric fluxes are defined consistently.

The uniform temperature (or equivalently, uniform geopotential) approximation has appeared in one guise or another in numerous treatments of the tropical general circulation (Lindzen et al. 1982; Schneider 1977; Lindzen and Nigam 1987; Wallace 1992; Neelin and Held 1987). Ramanathan and Collins also invoke it in their attempts to argue away the effects of cloud longwave forcing, though we shall see later that there are inconsistencies in their application of the principle. The simplest argument for the uniform temperature of the tropical atmosphere is that the local radius of deformation becomes large near the equator, where the Coriolis parameter vanishes. The essential ideas are implicit in the discussion of adjustment under gravity found in Gill (1982). This explanation is secure only within a few degrees of the equator, though. Further from the equator, there is still a powerful limit on meridional temperature gradients for *axisymmetric* motions, owing to the joint requirements of angular momentum conservation and the thermal wind balance (Hide 1969; Schneider 1977). The extent to which this constraint survives mild zonal asymmetry is still an unsettled question. Be that as it may, an appreciation of the validity of the uniform temperature assumption can be obtained by examining the 300-mb monthly mean temperature analyses shown in Fig. 5. In 1987, which is an El Niño year, the equatorial temperature varied by only 1 K around the entire planet. The La Niña year, 1989, exhibited somewhat greater variations, but these still amount to only 2 K along the equator. The Tropics are bounded by high temperature gradients, and the boundary is especially clear-cut in 1987. In 1989, the cold extratropical regime pinched into the central Pacific, leading to temperature variations of 3–4 K at 20°N and 20°S. The greater variability at the edge of the Tropics in 1989 is consistent with the breaking of Hide's theorem by the greater zonal asymmetries in that year. The 1987 – 1989 difference map shows that the entire Tropics warms up during the El Niño year, with the most pronounced warming occurring off the equator in the central Pacific. We conclude that the uniform temperature approximation is imperfect, especially toward the edge of the Tropics and especially in La Niña years. Still, the magnitude of the deviations does not seem so great as to compromise the utility of the assumption in a zero-order picture.

With the equilibrium and the efficient heat transport assumptions, we are now prepared to compute the sensitivity of the tropical temperature to changes in a control parameter, say λ. For example, λ could be the solar constant, or the atmospheric CO_2 concentration. If we wish to think about fluctuations in the course of the El Niño cycle, λ could be some measure of the strength of the tropical oceanic heat transport. Let T_r be some reference temperature characterizing the atmosphere. For our purposes, it is only necessary that the profile $T(\zeta)$ can be computed (e.g., using the saturated moist adiabat) once the single value T_r is given. The low-level air temperature is a natural choice of reference temperature for SST stability studies, but the tropopause temperature would serve equally well. According to the results of section 2, considering sensitivity to $T(0)$ is practically the same thing as considering sensitivity to warm pool SST. Now suppose that H can be written as a function of T_r and λ. The equilibrium temperature is then determined by finding T_r such that

$$H(T_r, \lambda) = 0. \tag{3.8a}$$

Now, if λ is changed by an amount $\Delta\lambda$, the temperature changes by an amount

$$\Delta T_r = - \frac{\left[\dfrac{\partial H}{\partial \lambda} \Delta\lambda\right]}{\beta}, \tag{3.8b}$$

where

$$\beta = \frac{\partial H}{\partial T_r}. \tag{3.8c}$$

The partial derivative in the numerator of (3.8b) is to be taken with T_r fixed, while that in (3.8c) is to be taken with λ fixed. Both derivatives are to be taken subject to the constraint of balanced surface energy budget. The numerator in (3.8b) gives the net heat flux perturbation due to a change in λ with T_r fixed and β is the sensitivity coefficient that determines how the atmospheric temperature responds to this shift. The existence of strongly stabilizing feedback mechanisms (i.e., large β) would lend to the tropical climate a highly interesting degree of resilience.

From (3.6) and (3.4) it follows that

$$\beta = - \frac{\partial}{\partial T_r} \langle I_c^+(\infty)\rangle + \frac{\partial}{\partial T_r} F_{aexp}$$

$$+ \frac{\partial}{\partial T_r} \langle C_s + C_l\rangle + \frac{\partial}{\partial T_r} \langle F_{oh}\rangle. \tag{3.9}$$

The final term in (3.9) represents feedback due to changes in the net oceanic heat flux out of the Tropics. The penultimate term in (3.9) is the cloud feedback term, which is negligible to the extent that $C_s + C_l = 0$. The second term in (3.9) is the feedback due to atmospheric heat export from the Tropics, and this

could indeed be a stabilizing feedback as it is entirely plausible that a warmer Tropics might export correspondingly more heat. However, because F_{aexp} may be as little as 20 W m^{-2} averaged over the region 20°N–20°S (Peixoto and Oort 1992), large fractional changes are required if it is to have a strong impact. This mechanism is eminently worth considering, but we do not know how to model the expected changes in F_{aexp} in any simple way, and the changes in any event most likely depend on the details of the extratropical circulation. Hence, though we believe F_{aexp} may be a significant feedback, we will treat it as an externally specified parameter of the tropical circulation rather than attempting to express it as a function of T_r. For similar reasons, we consider $\langle F_{oh}\rangle$ to be a specified parameter of the system rather than a feedback and set the oceanic term in (3.9) to zero by fiat.

Attention thus settles on the first term in (3.9), which gives the rate at which clear-sky OLR changes with increasing atmospheric temperature, subject to the constraint that the surface fluxes remain in equilibrium. It is the only surviving sensitivity factor that operates internally to the Tropics. A better understanding of the operation of this term can be obtained through examination of a few extreme limits.

First, suppose that the entire Tropics is saturated with respect to water vapor and remains so as temperature changes. In this case, the OLR increases only slightly with atmospheric temperature, as discussed in section 2. In consequence, the tropical climate will be very sensitive to perturbations. This is in essence the argument for amplification of climate sensitivity to doubling of CO_2, laid down by Manabe and Wetherald (1967). Now suppose that only a portion of the Tropics (which we shall think of as the "warm pool atmosphere") is saturated but that the rest of the tropical atmosphere (the "cold pool atmosphere") is completely dry and transparent in the infrared, as if it were composed of pure nitrogen. Suppose further that the sensible and latent heat exchanges with the cold pool ocean surface are negligible. In this case, the cold pool SST is independent of atmospheric temperature, whence, owing to transparency of the atmosphere, the cold pool OLR is also independent of atmospheric temperature. Thus, the sensitivity factor β is solely determined by the warm pool OLR curve and represents a sensitive climate as in the previous case. Because the cold pool atmosphere is radiatively passive in this case, the entire tropical atmosphere is controlled by the warm pool budget. Finally, suppose the cold pool atmosphere contains a *fixed concentration* of infrared emitters independent of temperature, as if its emissivity were solely due to CO_2. In that case, the OLR from the cold pool atmosphere rises roughly like the fourth power of the atmospheric temperature, even though the warm pool OLR is relatively flat with temperature. In this limit, the magnitude of β is increased and the cold pool has a stabilizing effect on tropical climate.
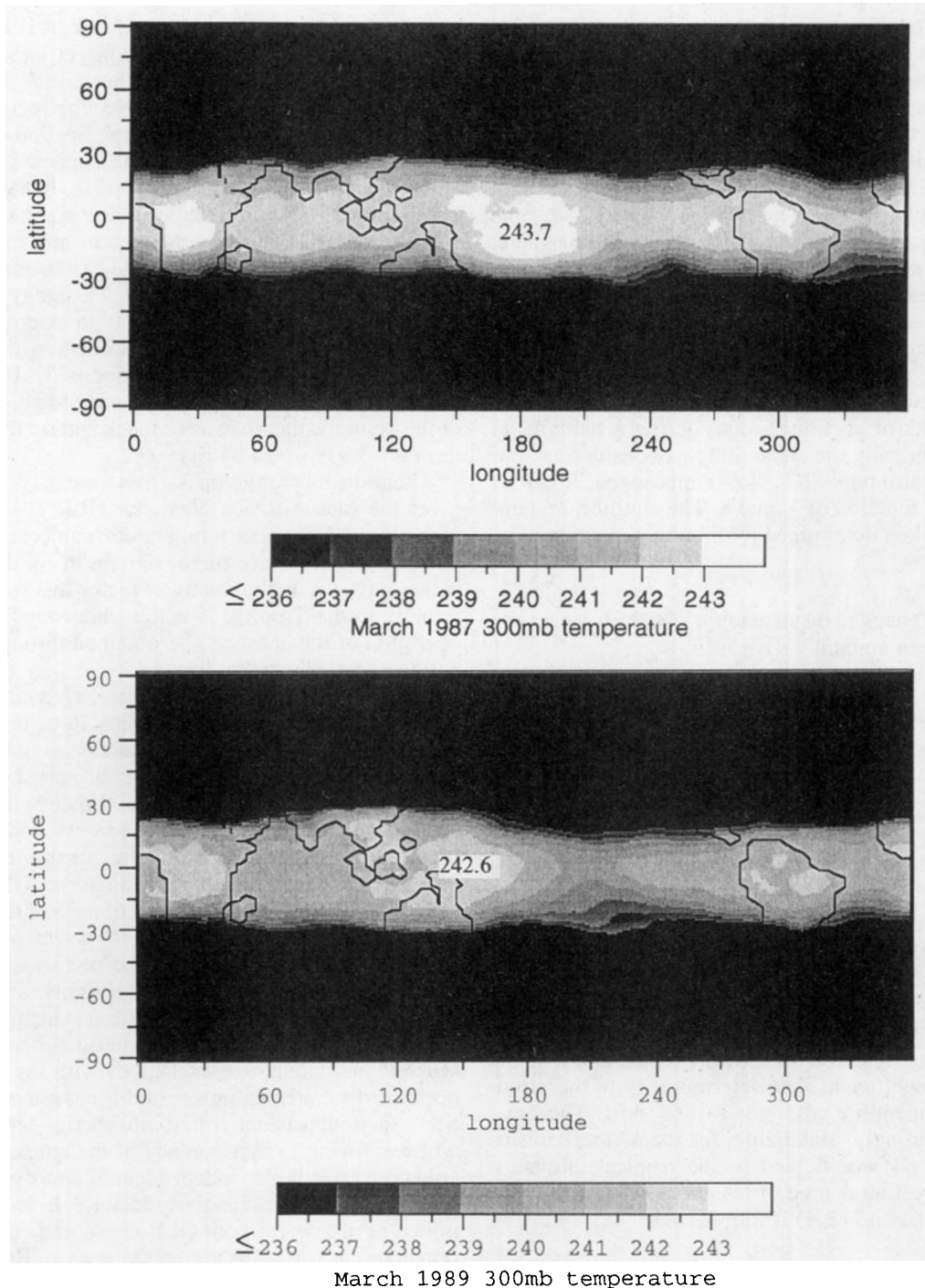
FIG. 5. Monthly-mean ECMWF 300-mb temperature analyses for March 1987, March 1989, and the difference of the two years. The dashed line in the difference map is the locus of zero temperature change.

The warm pool atmosphere cannot get rid of its heat, because of the strong water vapor greenhouse effect; this heat must be exported via zonal and meridional heat fluxes, to drier regions where it can be radiated to space. These dry, nonconvective regions act like "radiator fins" stuck into the side of the warm pool at-

mosphere. The "super greenhouse" shape of the clear-sky OLR curve in the analyses of Raval and Ramanathan (1989) and RC provides direct evidence for radiator fins, since it shows that OLR is generally higher in some cooler SST regions than it is over the warmest tropical waters. Hallberg and Inamdar (1993)
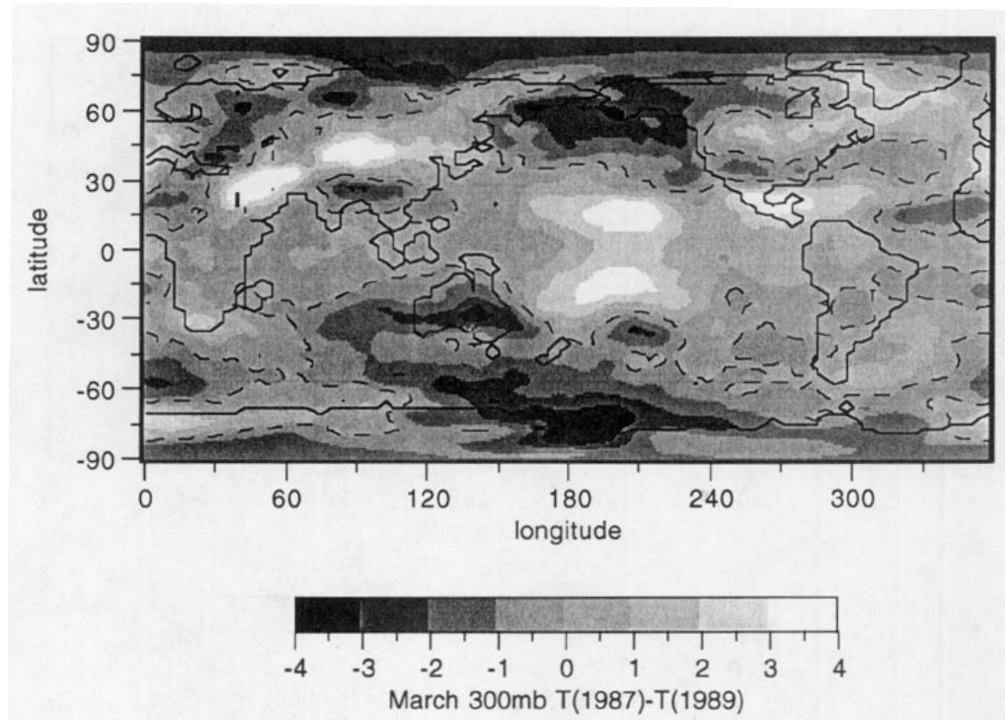
FIG. 5. (*Continued*)

showed that relative humidity variations are crucial to the observed drop in OLR with increasing SST; hence, the large-scale flow that modulates convection and controls the humidity distribution has an important effect on the tropical radiation budget. The lateral heat flux required to redistribute energy within the Tropics is carried largely by the Hadley and Walker circulations, which also provide the macroscopic subsidence that is responsible for dry, convectively suppressed regions. Hence, the atmosphere acts to maintain its own radiator fin. If this structure were somehow disrupted, the Tropics as a whole would receive more net heating and would have to either warm considerably or export more heat to the extratropics. Either would entail major changes in the global circulation.

Since the clear-sky emissivity is closely tied to water vapor (e.g., Stephens and Greenwald 1991), we can look at observed relative humidity patterns to get an idea of the disposition of radiating structures in the Tropics. In Fig. 6, we show the monthly-mean 700-mb ECMWF relative humidity analysis for March 1989 and March 1987. This "data" is actually mostly model created (Trenberth 1992), but satellite studies (Soden and Bretherton 1993, 1994) suggest that the ECMWF analyses capture the gross features of the moisture distribution. The La Niña year had a moist atmosphere in the western Pacific and a large dry area extending over the entire eastern tropical Pacific. The El Niño year, 1987, showed a thin, saturated band extending along the equator throughout much of the Pacific. The whole moisture pattern is more zonally sym-

metric than 1989, with pronounced dry bands in the subtropics corresponding to the descending branches of the Hadley circulation. The transition is indicative of a shift from radiating regions and heat transports dominated by the Walker circulation in normal years, to a greater dominance of the Hadley circulation in El Niño years.

We have not needed to use the detailed form of the surface energy budget (2.5), but a closer look at the terms in this budget provides some insight into why cloud effects appear neither in the atmospheric heating nor the expression for lateral heat transport. This may seem peculiar at first, because the cloud shortwave forcing primarily shows up as a cooling term at the surface, whereas the cloud greenhouse effect primarily shows up as an atmospheric warming. Because the net IR cooling of the surface is weak and the oceanic heat fluxes are of limited magnitude, the dominant balance in (2.5) is between absorbed solar radiation and evaporation (Philander 1990; Hartmann and Michelsen 1993; see also Fig. 6 in Washington and Meehl 1993). This implies that the primary effect of increasing cloudiness is to decrease the evaporation, an effect that is amply borne out in general circulation model simulations (Boer 1993; Washington and Meehl 1993). The reduction in evaporation as cloudiness increases reduces the latent heating of the atmosphere. This is the primary means of communicating the effect of C_s to the atmosphere, though it must be emphasized that none of the arguments leading to (3.9) actually rely on its dominance.
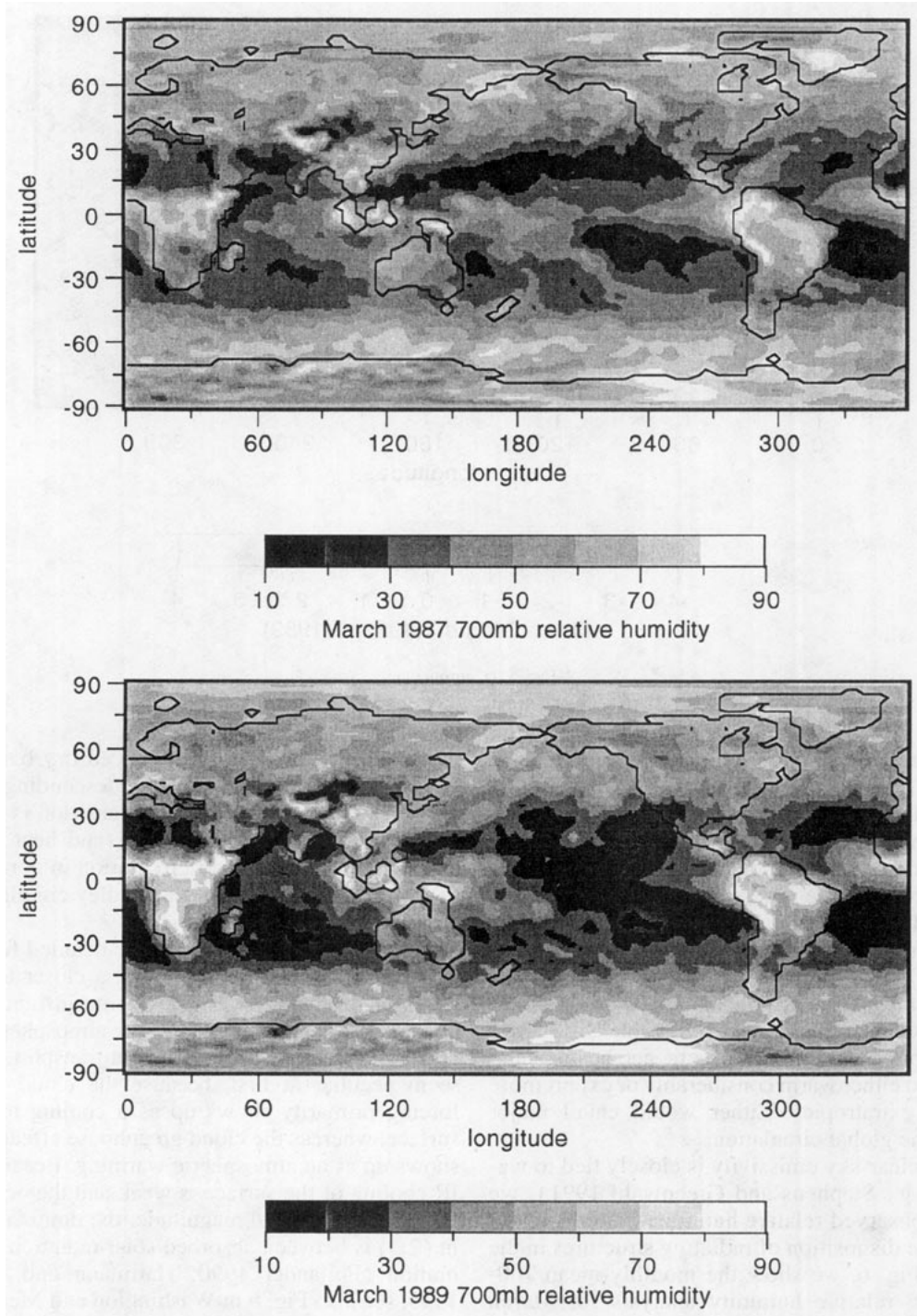
FIG. 6. Top panel: ECMWF monthly-mean 700-mb relative humidity for March 1987.
Bottom panel: the same for March 1989.

When interpreting general circulation model studies of cloud feedbacks in light of the above results, there are two variables to keep in mind: whether the study employs fixed sea surface temperatures and whether the modeled clouds conform to the observed constraint that $C_s + C_l \approx 0$ in the Tropics. Fixed SST simulations (e.g., Randall et al. 1989; Sherwood et al. 1994) are not energetically closed. They can be instructive, but the cloud effects cannot be taken as representative of what would happen if SST were allowed to adjust so

as to satisfy the surface energy balance. These calculations respond fully to C_l but see C_s only faintly through its effect on atmospheric absorption. Equilibrium climate studies with a mixed-layer ocean (Boer 1993; Washington and Meehl 1993) do respect the surface energy balance but do not necessarily maintain the observed approximate $C_s + C_l$ cancellation. There is no doubt that clouds would be a stabilizing influence if C_s were to become dominant; this possibility is explicitly present in (3.9). Boer et al. (1992) and Boer (1993) do not do a cloud–no-cloud comparison. Washington and Meehl (1993) do not diagnose $C_s + C_l$. Indeed, because they specify cloud albedo arbitrarily as a function of SST, while allowing the cloud longwave optical properties to be internally determined, it would be remarkable if the cancellation were found to be satisfied. To our knowledge, a cloud–no-cloud comparison has not yet been carried out in a model coupled to a mixed-layer ocean and in which the cloud scheme has been tuned to agree with satellite top-of-atmosphere observations.

4. A critique of the thermostat hypothesis

Arguing on the basis of essentially the same observational background as we do, RC come to very different conclusions regarding climate sensitivity than we have reached. They argue that the sensitivity coefficient β is dominated by the cloud shortwave forcing C_s and has magnitudes in the range of 15–26 W m^{-2} K^{-1}. With such a large sensitivity coefficient, RC find that the tropical SST would not exceed 305 K *even if the Tropics had to accommodate the entire imbalance between incoming and outgoing radiation at the top of the atmosphere*. This net imbalance amounts to 40–50 W m^{-2}, and RC's result would imply that the much smaller radiative forcing due to changing CO_2 would be wholly inconsequential. Quantitative estimates of β in the framework of section 3 will be given in the next section, but it is already clear from (3.9) that our sensitivity result is wholly different in that cloud effects largely drop out in favor of other effects. On examination, it is found that the differences between RC's assessment and our own come not so much from differences in the basic physical assumptions about tropical behavior as in the consistency with which those assumptions are applied.

First, we note that the thermostat hypothesis can be dismissed on quite general grounds, independent of the details of the calculations reported in the preceding sections. Ramanathan and Collins argue that the cloud albedo feedback increases sharply as the SST exceeds the threshold of 300 K at which deep convection sets in. What they fail to note is that the trigger temperature for deep convection is not a fixed constant for the planet. Deep convection becomes possible when the low-level air has enough moist static energy to reach the upper troposphere. Thus, if the troposphere warms,

the threshold temperature for convection and for the onset of cloud feedbacks also increases and, conversely, decreases if the troposphere cools. Hence, cloud feedbacks cannot control the absolute temperature of the Tropics; at best, they can affect the SST *differences* across the Tropics. This effect can be seen clearly in the GCM experiments of Washington and Meehl (1993). Their model is about 3 K warmer in the Tropics than the real climate, and deep convection as evidenced by the rise in the atmospheric greenhouse effect sets in at a higher temperature than in reality.[2] This is not to say that a uniformly warmer (or cooler) climate will be in every regard identical to the base climate. Of course, the differing moisture contents of the atmospheres will have numerous profound effects. Nonetheless, the threshold for onset of cloud feedbacks will vary closely with the overall warmth of the atmosphere.

Actually, having shown that the thermostat hypothesis cannot control overall tropical temperature, there is not much content left in the hypothesis at all. Without question, the atmospheric temperature is kept horizontally uniform in the Tropics by essentially dynamical means. Since, for the reasons noted in section 2, the sea surface temperature cannot much exceed the low-level air temperature, it is guaranteed that SST will saturate at a value close to this uniform temperature. On the other hand, where there is sufficient cooling of the ocean mixed layer, the SST is free to drop without any a priori bound. This is essentially the argument put forth by Wallace (1992).

The specific disagreement between our formulation and that of RC comes in the estimate of the lateral atmospheric heat transports within the Tropics. Whereas we compute the heat transports by finding the flux needed to keep the tropical atmospheric temperature uniform, RC rather arbitrarily set $F_{ah} = (1 - f)C_l$, where f is a small factor purporting to represent the effectiveness of the atmospheric heat redistribution. Thus, RC's assumption requires the atmospheric dynamics to selectively redistribute the heat due to cloud longwave forcing, while ignoring all other forms of atmospheric heating—notably latent heating. We leave it to the reader to judge whether there are any circumstances in which this is a plausible assumption. The estimate by RC also fails to recognize that the atmospheric heat transport responds not to the warm pool atmospheric heating itself but to the *contrasts* in atmospheric heating across the Tropics.

It should also be noted that RC refer to an "f factor" in an additional context quite distinct from that appearing in their lateral heat transport estimate. This sec-

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[2] Note that while Washington and Meehl impose the temperature dependence of cloud albedo feedback by fiat, the convection and cloud longwave forcing are determined internally by the model dynamics.

ond sense of the f factor, call it f', is the ratio of the cloud-induced change in OLR to the change in downwelling infrared received by the surface. This factor, f', is amenable to measurement through comparison of surface and top-of-atmosphere observations, but experimental verification of small f' in no way supports the thermostat hypothesis. Small f' means only that the cloud longwave heating shows up mostly as an atmospheric rather than a surface heating; it says nothing about how the atmosphere redistributes this heat.

In RC's picture of the operation of the tropical climate, the warm pool is in a runaway state, and the only brake is provided by clouds. They imply further that the induced atmospheric circulation brings in more moisture, which only further destabilizes the system, somewhat after the fashion of wave–CISK. We have already seen in sections 2 and 3 that the tropical clouds cannot stabilize the runaway greenhouse. In the next section, we will see that in fact an equable tropical climate can be maintained without invoking clouds. It will also be seen that moisture import into the warm pool region does not generally render the coupled system unstable.

5. A two-box model of the equilibrium tropical circulation

In this section, we will show how the considerations of section 3 play out within a quantitative, though highly idealized, model of the tropical general circulation. We retain all the key assumptions of section 3, notably the equilibrium and efficient heat transport assumptions, but strip the system down to its basics by reducing the spatial structure of the Tropics to only two boxes, shown schematically in Fig. 7. In line with the efficient heat transport assumption, the atmospheric temperature profile $T(\zeta)$ in the two boxes is the same; however, the emissivities, latent heating, and sea surface temperature are allowed to be different between the boxes. The first box (the "furnace" or warm pool atmosphere) is thought of as a region of net ascending motion and deep convection, with consequent latent heating of the atmosphere. Because of the moisture source from the deep convection, we presume this region to have rather high relative humidity, whence the atmosphere is optically thick in the infrared. The second box (the radiator fin or cold pool atmosphere) is a region of mean descent. There is no deep convection or latent heat release in the interior of the atmosphere so that the diabatic effects there are purely radiative. The adiabatic warming due to the dry descent must be balanced by a net diabatic cooling in this region. In accord with the mean descent and the lack of deep convection, it is presumed that the radiator fin atmosphere is relatively dry above the boundary layer and hence that the free atmosphere is not very optically thick in the infrared. Evaporation is allowed from the cold pool surface, but the moisture is presumed to be transported

to the warm pool, where it is released as deep latent heating. The circulation sketched in Fig. 7 is to be thought of as a proxy for the joint effects of the Hadley and Walker circulations.

The notation, and terms in the budgets, are identical to sections 2 and 3, save that we append a subscript 1 or 2 to designate the box to which the term applies. Note that in Fig. 7 we have highlighted only a few of the flux terms entering into the budgets. If we take the areas of the two boxes to be A_1 and A_2, (3.6) becomes

$$(Q_{v1} + F_{a\exp}) \frac{A_1}{A_1 + A_2} + (Q_{v2} + F_{a\exp}) \frac{A_2}{A_1 + A_2} = 0$$
(5.1)

in equilibrium. The lateral heat transports are mediated by mass fluxes, which cause adiabatic compressional warming or cooling within each column. Let M_1 be the vertical mass flux (mass per unit area per unit time, upward positive) in the furnace region, and M_2 be the mass flux in the radiator fin region. The flux M_2 is determined by balancing the adiabatic warming by dry descent against the net diabatic effect in the radiator fin region. Assuming the cooling from mixing with cold extratropical air to be uniformly distributed across the Tropics, the net diabatic effect is then $Q_2 + F_{a\exp}$. Defined according to (3.7a), Q_2 is purely radiative because there is no precipitation in the free troposphere of the radiator fin region. Consistency with the presumption of subsidence requires $Q_2 + F_{a\exp} < 0$, that is, that it be a cooling. The balance implies

$$M_2 = \frac{Q_2 + F_{a\exp}}{C_p \delta\theta},$$
(5.2)

where C_p is the specific heat at constant pressure, and $\delta\theta$ is the difference between the dry potential temperature at the inflow to the top of the subsiding region and the outflow at its bottom. The flux M_1 can then be determined by closing the mass budget in the Tropics, but we shall not need it explicitly.

What we have outlined is a type of model of the tropical climate that is sometimes referred to as a "single-cell model." Other treatments of the Tropics in this framework include Schneider (1977), Sarachik (1978), Satoh and Hayashi (1992), Lindzen et al. (1982), Sun and Lindzen (1993a), and Betts and Ridgway (1989). Single-cell models come in many flavors, differing in the nature of the mass flux closure, the details of radiative transfer, the parameterization of convection, and the surface flux parameterizations. The discussion up to this point is applicable to virtually any single-cell model of the tropical circulation.

a. Assumptions peculiar to the present model

Our version of the single-cell model is distinguished primarily by a choice of some radical simplifications that allow us to bring out the essential behavior trans-
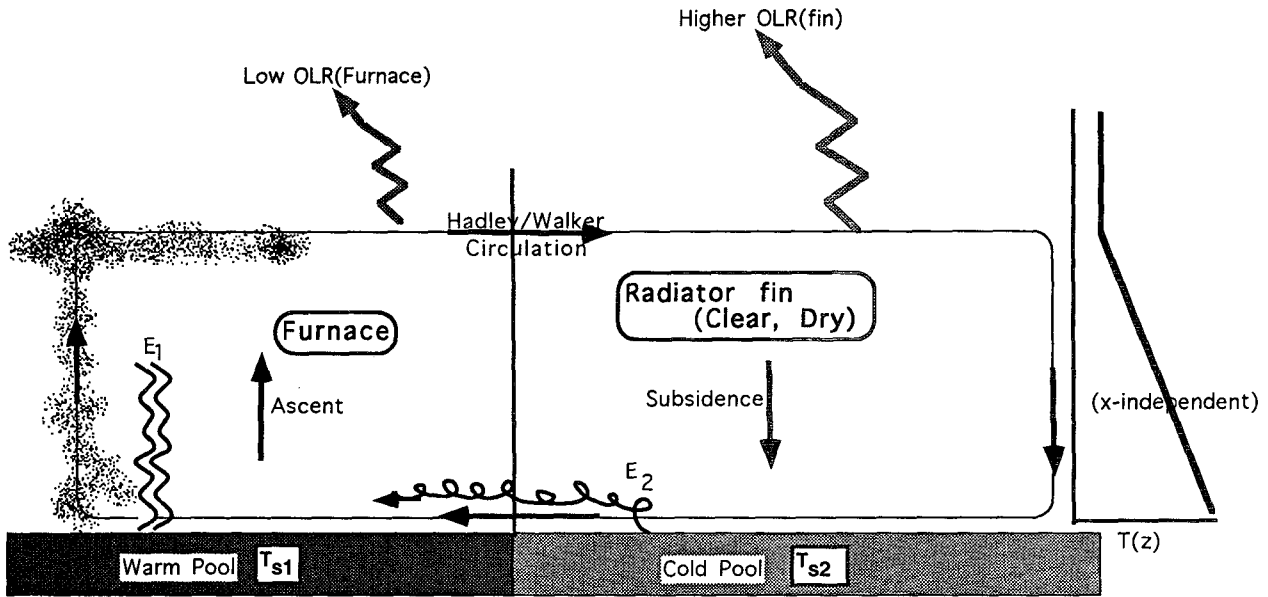
FIG. 7. Schematic representation of the furnace–radiator-fin model of the tropical general circulation.

parently. The chief utility of the model is didactic. We introduce it to bring out in concrete terms the repercussions of some of the phenomena discussed in general terms in section 3. It has too many adjustable parameters and too much missing physics to enable reliable quantitative projections of climate change to be made, but it will nonetheless be of interest to see whether such a model can be made to yield earthlike conditions. A more substantive distinctive feature of our variant is that we consider the ascending, moist region to have finite area, so that it participates directly in the radiation budget of the Tropics. We also compute the SST separately in the warm pool and the cold pool regions. The other single-cell models cited essentially take the limit of small A_1 with M_1A_1 fixed. This is done on the indisputable grounds that the actual area occupied by ascending cumulus turrets constitutes a very small portion of the Tropics. We take the viewpoint that the ensemble of ascending turrets in the large-scale convective region over the warm pool succeed in moistening a sufficiently large area of their environment that it is plausible to define a macroscopic-moistened region. In our picture, M_1 represents the residual large-scale upward mass flux in this region rather than the upward flux in the individual turrets. No doubt, there is also smaller-scale subsidence embedded within the warm pool atmosphere, which constitutes an important drying mechanism to be balanced against the moisture source from detrained and precipitated cloud water. The proximity of these microsubsiding regions to the moisture source nonetheless permits the warm pool atmosphere to become moist. Support for this picture comes from the ECMWF moisture analyses of Fig. 6, from the tropical moisture soundings discussed in In-

amdar and Ramanathan (1994), and more directly from the clear-sky OLR behavior in Raval and Ramanathan (1989) and Stephens and Greenwald (1991). Schneider (1977) and Lindzen et al. (1982) allow geographical variations in the surface temperature but do not allow for large-scale modulation of the humidity field in the same sense that we do.

Now we enumerate the specific assumptions defining our particular variant of the single-cell model, to wit:

(i) We employ the fixed temperature profile (2.4), adjusting only $T(0)$. Thus, we neglect the effect of lapse rate and tropopause height feedbacks, in favor of focusing on other aspects of tropical climate stability.

(ii) The warm pool is considered to have fixed relative humidity and to be optically thick in the infrared. The value of Q_{v1} is obtained directly from (3.4), with I_c^+ obtained as a function of $T(0)$ from Fig. 2. Assumption (i) affects the detailed shape of the warm pool OLR curve. Note further that solar absorption in the atmosphere is allowed for implicitly, though it does not appear explicitly in the expression for Q_{v1}.

(iii) As suggested by the discussion of surface energy budgets in section 2, we simply set $T_{s1} = T(0)$ + 1 K and do not bother with a detailed solution of the surface temperature equation in the warm pool. Because the warm pool SST does not much affect OLR, an improved treatment of the surface energy budget could always be incorporated after the fact by simply inserting the appropriate value of $[T_{1s} - T(0)]$ into the final results presented below.

(iv) The radiator fin atmosphere is assumed to radiate as an equivalent greybody with emissivity e_2 and effective temperature $\bar{T} = T(\zeta_{rad2})$, where ζ_{rad2} is a

specified radiating level. Solar absorption is allowed for by means of the coefficient α introduced in (2.5). There are no clouds in the cold pool region.

(v) The cold pool evaporation is taken to be controlled by the mass flux M_2, which delivers dry air to the boundary layer. This is similar to the assumption employed by Betts and Ridgway (1989). If q_2^* is the characteristic water mixing ratio of the air transported from the cold pool region into the warm pool region, then the cold pool evaporative heat flux is thus

$$E_2 = -Lq_2^*M_2 = -b(Q_2 + F_{aexp}), \quad (5.3a)$$

where

$$b = \frac{Lq_2^*}{C_p\delta\theta}, \quad (5.3b)$$

and the sign convention is that positive E_2 represents a cooling of the cold pool ocean surface. The moisture mixing ratio of the cold pool outflow is estimated by

$$q_2^* = r_2 q_{sat}(T_{s2}), \quad (5.3c)$$

in which r_2 is a specified coefficient characterizing the efficiency with which the cold pool boundary layer moistens the low-level return current of the Hadley–Walker circulation. If the return current were confined to the boundary layer, r_2 would be the relative humidity of the cold pool boundary layer air. While this provides an upper bound for r_2, the true value is apt to be rather smaller, owing to dilution of boundary-layer air with drier air from above. Hartmann and Michelsen (1993) have formulated a model of the low-level tropical flow, which provides more insight into the physics determining moisture export from the cold pool.

(vi) The cold pool SST is determined by the surface energy budget (2.5), with $C_s^* = 0$ and formulas for E_2 and $I^-(0)$ obtained from the above assumptions. We neglect sensible heat transfer, but allow for the oceanic heat flux F_{oh2} as a specified parameter.

(vii) We allow the warm pool to be cloudy but assume that $C_s + C_l = 0$. With assumption (iii) concerning the warm pool SST, clouds simply act to reduce the warm pool evaporation. This effect has already been adequately discussed above and we will not dwell on it further in this section.

(viii) The thermodynamic depth $\delta\theta$ of Hadley/Walker circulation is governed by complex dynamical and thermodynamic effects that are not easy to faithfully incorporate into the idealized framework treated here. We have opted for the expediency of a *fixed* $\delta\theta$, which at least has the virtue of making the results easy to interpret.

The basic set of parameters we employ are $h = 7$ km, $\gamma = 6.2$ K km^{-1}, $\zeta_T = 16$ km, $S_c = 370$ W m^{-2}, $\alpha = 0.8$, and $\zeta_{rad2} = 1/2h$. The last of these parameters is the most arbitrary, and we have chosen it partly by tuning the model so that it produces earthlike temper-

atures when allowance is made for heat fluxes out of the Tropics. The choice $\zeta_{rad2} = h$ (rigorously appropriate to an optically thin atmosphere) was found to yield unrealistically warm climates. A lower radiating level may be rationalized by presuming that there is some moistening of the lower part of the radiator fin troposphere. Except where otherwise noted, we used $r_2 = 70\%$ and $\delta\theta = 85$ K, which are roughly the largest values that could conceivably be appropriate to the present climate. These enter the problem only in the ratio $\delta\theta/r_2$, so a choice of a smaller thermodynamic depth and proportionately drier low-level outflow would give the same results. Other control parameters (warm pool relative humidity, F_{aexp}, and F_{oh2}) are varied as noted below.

The fact that longwave radiation into the sea surface is primarily determined by low-level moisture would appear to be a fatal flaw in assumption (iv), since $I^-(0)$ would remain rather close to $\sigma T(0)^4$ even if only the boundary layer had high humidity. A simple sleight of hand involving shallow convection and the trade boundary layer makes most of this objection disappear. We think of T_{s2} as characterizing the temperature of the coupled ocean surface–trade boundary layer system rather than just the ocean surface per se. Shallow convection and precipitation are allowed within the boundary layer, which serves to keep the SST similar to the temperature of the air with which it is immediately in contact. This is admittedly an extremely crude representation of the trade boundary layer; the reader is referred to Betts and Ridgway (1989, 1992) for a much more realistic and illuminating discussion of the tropical boundary layer. In our picture, e_2 is to be thought of as the effective emissivity of the free, dry, subsiding atmosphere above the trade inversion. Likewise, E_2, computed according to assumption (v), does not represent the full evaporation from the cold pool sea surface but only that portion that escapes the cold pool and is transported to the warm pool deep convection region. The remaining evaporation (if any) needed to keep down the cold pool SST is presumably returned to the local ocean in the form of rain from shallow precipitating systems within the tropical boundary layer.

The most worrisome aspect of our formulation is the assumption of a fixed temperature profile. In particular, one might expect the lapse rate to somewhat follow the moist adiabat and therefore go down in a warmer atmosphere. Note that the fixed-profile assumption enters in an essential way only in the determination of the warm pool OLR curves. However, most of our qualitative results depend only on the existence of a ceiling to the OLR radiated by a moist atmosphere, and as discussed in section 2, these limits are rather robust against changes of the shape of the temperature profile. The lapse rate also has some bearing on the processes determining $\delta\theta$, which together with r_2, is fixed rather arbitrarily in the present model. Both effects show up

only through the agency of b. Thus, we can be on the alert for junctures where the more dubious assumptions strongly affect the results by keeping a wary eye on the behavior of b.

b. Basic behavior of the model

Given the above assumptions, the diabatic heating/cooling in the radiator fin is purely radiative and is given by

$$Q_2 = (1 - \alpha)S_c + e_2\sigma T_{2s}^4 - 2e_2\sigma \bar{T}^4. \quad (5.4)$$

With this expression and (5.3a), the evaporation can be eliminated from the cold pool surface budget to yield

$$\sigma T_{2s}^4 = \frac{1}{1 - e_2 b}[(\alpha + (1 - \alpha)b)S_c$$

$$+ (1 - 2b)e_2\sigma \bar{T}^4 + bF_{aexp} + F_{oh2}]. \quad (5.5)$$

With this result in hand, the radiator fin OLR can be readily calculated, whence

$$Q_{v2} + F_{aexp} = \frac{1 - b}{1 - e_2 b}[(1 + (e_2 - 1)\alpha)S_c$$

$$+ e_2(e_2 - 2)\sigma \bar{T}^4 + F_{aexp} + e_2 F_{oh2}]. \quad (5.6)$$

This is not quite a closed form solution, because b depends on T_{s2} through q_2^*. An iteration on q_2^* is carried out at the final stage to meet the target relative humidity r_2.

Consistency of the solution requires $Q_2 + F_{aexp} < 0$. Since there is no precipitation in the free radiator fin atmosphere, (3.7a) and (5.3a) imply $(1 - b)(Q_2 + F_{aexp}) = (Q_{v2} + F_{aexp})$. On the other hand, for the radiator fin to help the warm pool atmosphere to get rid of its heat, we must have $Q_{v2} + F_{aexp} < 0$. Hence, the whole radiator fin structure breaks down if $b > 1$. Thus, b emerges as a key characteristic of the tropical circulation. If $b > 1$, the atmosphere may try to get rid of the excess warm pool heat by creating a large-scale circulation, but the latent heating imported by the circulation exceeds the sensible heat exported—in other words, the tropical circulation becomes thermally indirect. Thus, b plays the same role as the "effective moist stability" introduced by Neelin and Held (1987) in their model of the tropical circulation. Because we hold $\delta\theta$ fixed, b is purely a function of T_{s2}; with $\delta\theta = 85$ K and $r_2 = 0.7$, we have $b(280$ K$) = 0.13, b(290$ K$) = 0.25, b(295$ K$) = 0.34, b(300$ K$) = 0.47, b(305$ K$) = 0.63, b(310$ K$) = 0.84$, and the breakdown would occur at $T_{s2} = 313$ K. However, the breakdown is generally prevented from occurring by a stabilizing feedback inherent in (5.5). When $b > 1/2$, T_{s2} *decreases* with increasing $\bar{T}$, because increases in cold pool evaporation outweigh the increases in cold pool radiative heating of the surface. When b starts to get too close to unity, $\bar{T}$ becomes large because the warm pool is more nearly on its own, but then T_{s2} decreases, which

would reduce b. As a result, we have consistently been able to find equilibrium solutions with $b < 1$, and the hypothetical breakdown is never realized.

There are any number of ways to organize the solution of our two-box model, but a treatment in terms of the warm pool budget is the most illuminating. Assume for the moment that $F_{aexp} = F_{oh1} = F_{oh2} = 0$. From (3.5), F_{ah1} reduces to $(A_2/A_1)Q_{v2}$ for the two-box model. To determine the climate, we now balance $S_c + F_{ah1}$ against the OLR, instead of just balancing S_c, as would be the case for an isolated column. According to (5.6), F_{ah1} can be made as negative as desired by making $\bar{T}$ large, provided that $e_2 > 0$, $b < 1$, and b does not approach unity. Hence, even though the OLR may be flat as a function of $T(0)$, the heat flux F_{ah1} can supply sufficient cooling to stabilize the runaway greenhouse at finite $\bar{T}$. A graphical determination of the temperature by balancing $S_c + F_{ah1}$ against warm pool OLR is shown in Fig. 8, with 100% warm pool relative humidity and q_2^* held fixed. The stability of the climate to perturbations is determined by the angle of crossing of the two curves. If the curves cross at a steep angle, a small perturbation to the heat budget via either term leads to only a small shift of the intersection point. If the curves are nearly parallel near the crossing, a small change in conditions leads to a large change in temperature. Since the warm pool OLR curve is flat, the climate stability is provided mainly by the steep downward slope of F_{ah1}. This slope is directly traced to the *fixed emissivity* of the radiator fin atmosphere. The real atmosphere would behave this way if the moisture *concentration* rather than the *relative humidity* were invariant in the subsiding regions.

From (5.6) and the expression for F_{ah1}, we can infer the following general influences on the climate, supposing $b < 1$. The limit of small A_2/A_1 is singular, since no matter how small it may be F_{ah1} is negative for suf-
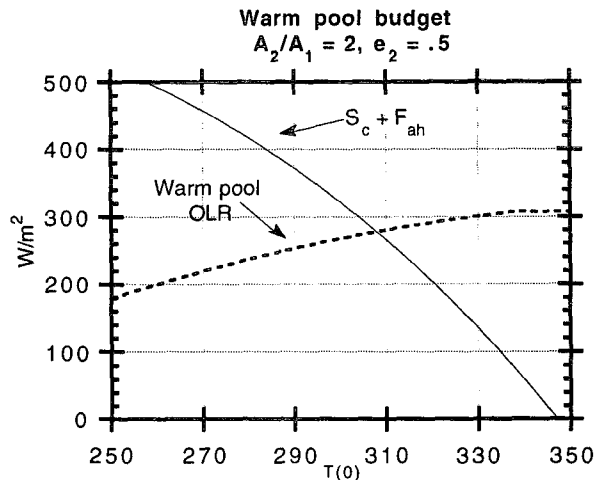

FIG. 8. Determination of atmospheric temperature via the energy budget of the warm pool atmosphere.

ficiently large $\bar{T}$, and the cooling of the warm pool atmosphere can be made arbitrarily strong by making $\bar{T}$ large, so long as A_2/A_1 is not zero. Increasing A_2/A_1 increases the cooling for fixed $\bar{T}$ and thus generally leads to a cooler climate. The limit of large A_2 is regular because Q_{v1} is bounded by the OLR limits for a moist atmosphere discussed in section 2. In this limit the temperature is determined by the cold pool box alone via $Q_{v2} + F_{aexp} = 0$. Both $\bar{T}$ and T_{s2} attain finite limiting values that are independent of the value of b. Returning to the general case, making F_{oh2} more negative (i.e., increasing cooling by oceanic heat fluxes in the cold pool) cools the climate, but the impact is proportional to e_2 since the effect is communicated to the atmosphere through IR absorption in the radiator fin. Increasing e_2 also cools the climate, provided $\bar{T}$ is large enough to overcome the influence of atmospheric solar absorption. Finally, the cooling becomes ineffective as $b \rightarrow 1$ with fixed $e_2 < 1$, owing to the feedback of imported latent heating. The temperature of the atmosphere approaches that of an isolated warm pool in this limit. However, when $e_2 = 1$, the cooling (and hence $\bar{T}$) becomes independent of b because the radiative effects implicit in the reduction of T_{s2} with increasing b exactly offset the changes in latent heat import.

c. Numerical results on parameter dependence

In Fig. 9, we show warm pool and cold pool SST as a function of e_2 for various A_2/A_1. This calculation was carried out with the Tropics isolated from the rest of the planet—that is, $F_{aexp} = F_{oh1} = F_{oh2} = 0$—and with 100% warm pool relative humidity. Note that for small e_2, corresponding to a very low CO_2 world with a very dry radiator, the tropical temperature runs away to very large values. A nonemissive "fin" does not make a good radiator. Ironically, increasing e_2 cools the climate. The temperature drops sharply as e_2 is increased to around 0.25; further increases in e_2 have a lesser effect on temperature. There is a broad minimum temperature at intermediate values of e_2, with a gentle rise as e_2 approaches unity. Note also that the furnace–radiator fin structure maintains cold pool SSTs lower than warm pool SSTs without the benefit of any oceanic circulations. Thus, cooling by oceanic cold upwelling may mainly serve to lock in a pattern that the atmosphere–ocean system is already predisposed toward on radiative grounds. Increasing e_2 brings the cold pool and warm pool temperatures closer together. In fact, for $A_2/A_1 > 2$, T_{s2} slightly exceeds T_{s1} for sufficiently large e_2. This happens because the cold pool comes to control the atmospheric temperature, and there is not enough evaporation to keep down T_{s2} in the face of the combined insolation and downwelling IR.

Even without heat exports from the Tropics, the system can maintain habitable temperatures given sufficiently large A_2/A_1. For $A_2/A_1 = 3$ and $e_2 = 0.55$ we find temperatures as low as $T_{s1} = 304$ K and $T_{s2} = 300$

K, which are only slightly warmer than those observed in the present tropical oceans.

In Fig. 10, we turn to a somewhat more realistic case, with warm pool relative humidity reduced to 75%. We took $F_{aexp} = -20$ W m^{-2}, as estimated in round numbers from Peixoto and Oort (1992). We chose $F_{oh2} = -30$ W m^{-2}, based on the zonal-mean heat flux across 20°N–20°S from the same source. For simplicity, we kept $F_{oh1} = 0$, which means that the net oceanic heat flux is somewhat underestimated in comparison to data. The same general behavior found in the previous case is seen, save that the climates are generally cooler. The onset of the situation where T_{s2} exceeds T_{s1} occurs at smaller A_2 than in the previous case, because the runaway greenhouse is not as effective in heating the atmosphere, and F_{aexp} further cools the atmosphere relative to the underlying surface. Also, the amount by which T_{s2} exceeds T_{s1} becomes larger. Since the warm pool SST is $T(0) + 1$ K, T_{s2} greatly exceeds the overlying air temperature $T(0)$. The cold pool atmosphere becomes very unstable to deep convection in this circumstance. Thus, when e_2 is not too small, there is a significant limit to how large the relative area of the cold pool can become. It is reasonable to assume that if the limit were exceeded, part of the cold pool would rapidly become convective, bringing the area back down to within the allowable limits. A realistic climate can be obtained with $A_2/A_1 = 1.25$ and $e_2 = 0.66$, which yields $T_{s1} = 301$ K and $T_{s2} = 298$ K.

An indication of the sensitivity of the results to $\delta\theta/r_2$ can be obtained from Fig. 11, where we show T_{s1}, T_{s2}, and b as functions of $\delta\theta$ (with $r_2 = 0.7$) for $A_2/A_1 = 3$, $e_2 = 0.7$, and the other parameters as for Fig. 10. Reducing $\delta\theta$ causes b to come closer to unity, which increases the warm pool temperature and, owing to greater evaporation, cools the cold pool. For sufficiently small $\delta\theta$, T_{s2} becomes smaller than T_{s1}, which is not possible at such large A_2/A_1 in the base case of Fig. 10. Thus, reduction of $\delta\theta$ allows the area of the cold pool to increase without going unstable, though at the expense of increasing the warm pool temperature somewhat. Now suppose that $\delta\theta$ is made proportional to the tropospheric dry potential temperature range for the saturated moist adiabat, instead of being kept fixed as in Fig. 10. This would be a stabilizing effect on the warm side, in the sense that a warmer atmosphere yields larger $\delta\theta$, which reduces b, increases efficiency of lateral heat transport, and thus limits the warming. However, it also tends to reduce the maximum value of A_2/A_1, for which the cold pool remains stable. Conversely, the $\delta\theta$ feedback would amplify sensitivity of the climate to cooling influences but allow larger cold pool areas.

d. Sensitivity of the climate

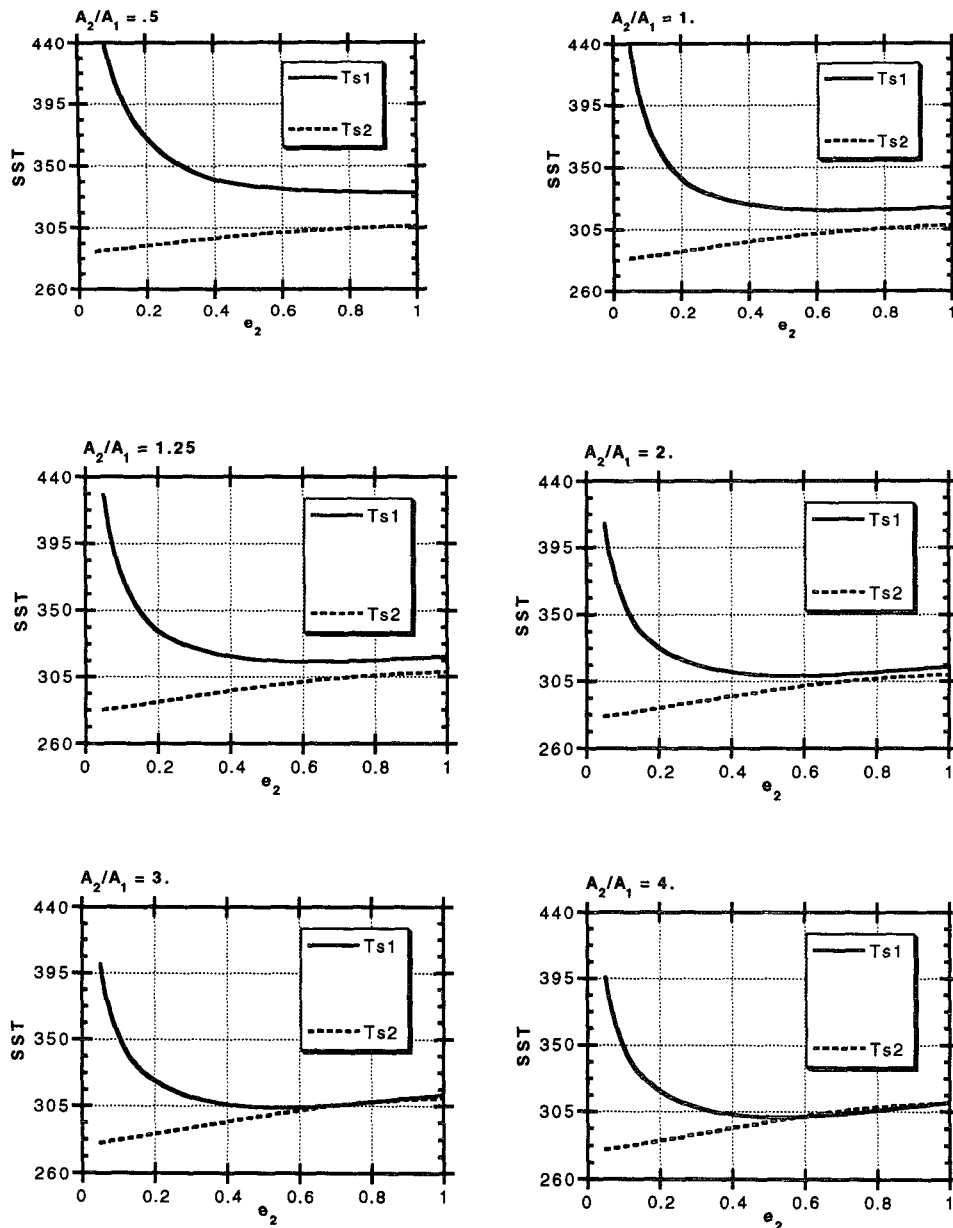In Fig. 12, we show some climate sensitivity results for $A_2/A_1 = 1.25$, with other parameters as for Fig. 10.

FIG. 9. Warm pool (T_{s1}) and cold pool (T_{s2}) SST as a function of e_2 for various values of A_2/A_1. Calculation carried out with 100% relative humidity in the warm pool atmosphere and an isolated Tropics $(F_{aexp} = 0,$ $F_{oh1} = F_{oh2} = 0)$.

We show the sensitivity coefficient β as defined by (3.8c) and also the net sensitivity factor $-\Delta S_c/\Delta T_{s1}$ for changes in insolation. These results are not to be taken too seriously, as (apart from the missing aspects of the physics) they represent sensitivity with A_2/A_1 and e_2 held fixed, whereas climate change could easily involve alterations in these parameters. On the other hand, though our model is crude, it is perhaps no cruder than the representation of moisture and convection within current general circulation models. Note that

$\Delta S_c/\Delta T_{s1}$ is somewhat larger than $|\beta|$, conferring a bit more stability than one would expect from β. One might have thought, from (3.8b), (2.1), and (2.3) that the two would be the same. In fact, they differ because $\partial H/\partial S_c < 1$. This partial derivative is taken with the atmospheric temperature fixed, but with the surface temperature kept in equilibrium. Thus, as S_c is increased, T_{s2} increases, which increases the cold pool OLR owing to the partial transparency of the radiator fin atmosphere. As expected from this argument, the
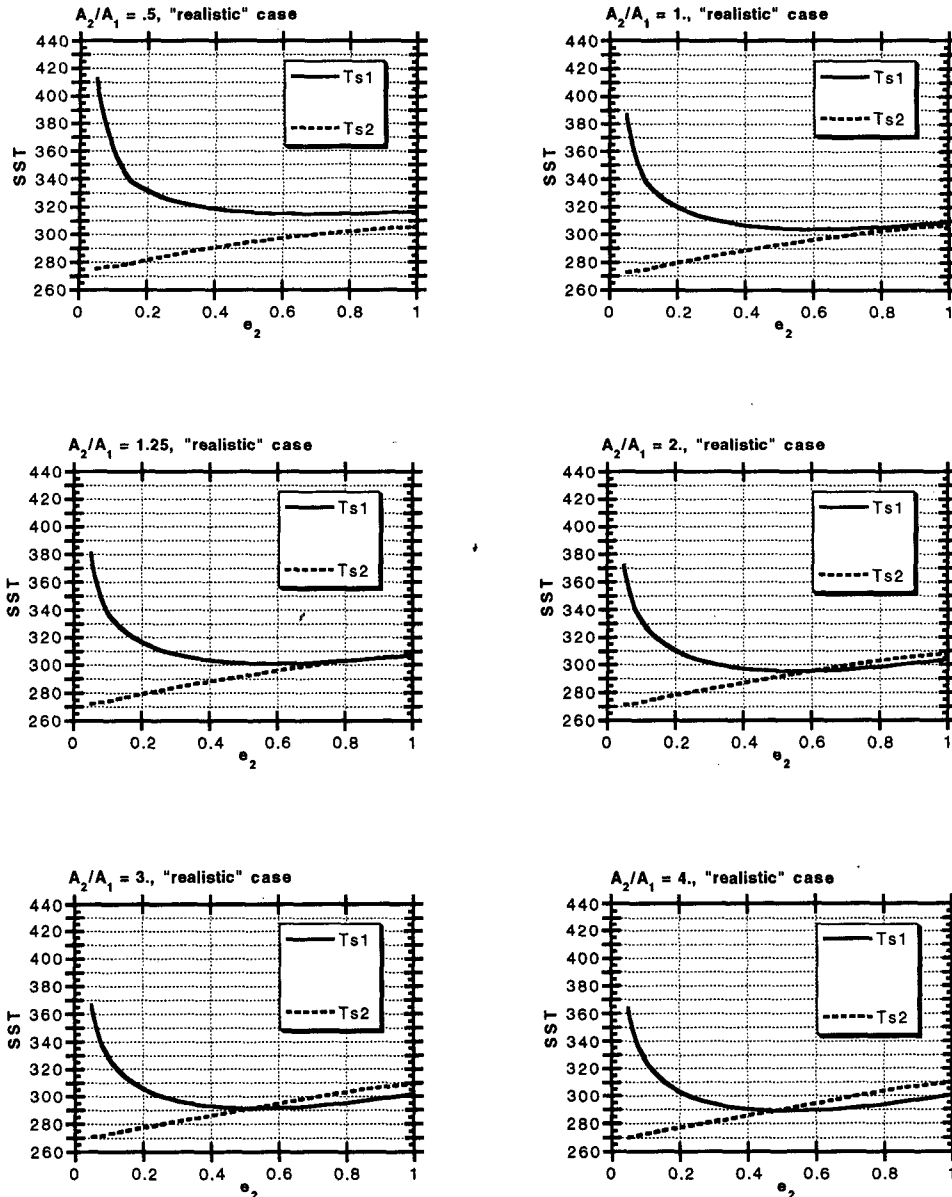
FIG. 10. As for Fig. 9 but for a "realistic" case with 75% warm pool relative humidity, $F_{oh2} = -30$ W m^{-2}, and $F_{aexp} = -20$ W m^{-2}.

two sensitivity factors converge on each other as $e_2 \to 1$. Sensitivity measured by either coefficient decreases monotonically with increasing e_2, as the effects of the fixed-emissivity region come to dominate the influence of the flat OLR curves of the fixed relative humidity region.

To provide a clearer idea of how the system accommodates changes in radiative forcing, we depict in Fig. 13 the changes in the warm pool energy budget caused by changing S_c by 10 W m^{-2}. This figure is for $A_2/A_1 = 1.25$, $e_2 = 0.6$, and other parameters as for Fig. 10. The warming of the atmosphere

increases the warm pool OLR by 5 W m^{-2}. The increased strength of the Hadley/Walker circulation yields a 16 W m^{-2} increase in the sensible heat export to the cold pool, but fully 11 W m^{-2} of this is cancelled by the concomitant increase in latent heat import from the cold pool. The bottom line is a 3.4 K warming of the warm pool SST. If the atmosphere had to get rid of the additional insolation without the aid of the increased lateral heat fluxes, it would warm by 8.8 K instead, as estimated from the slope of the OLR curve near $T(0) = 300$ K, for 75% relative humidity.
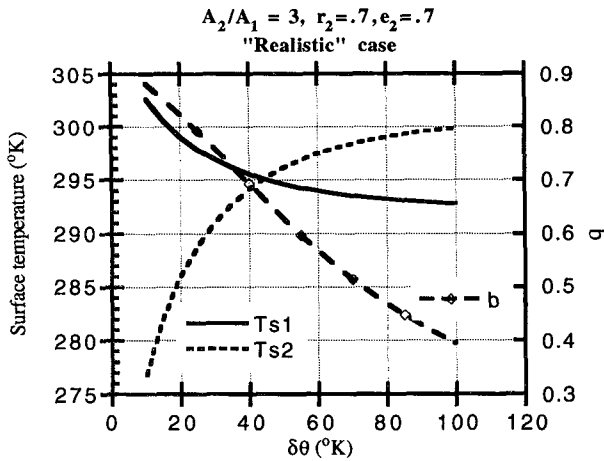
FIG. 11. Surface temperatures and moist stability parameter b, as a function of $d\theta$. Calculated with 75% warm pool relative humidity, $F_{oh2} = -30$ W m^{-2}, and $F_{aexp} = -20$ W m^{-2}, and other parameters as indicated.
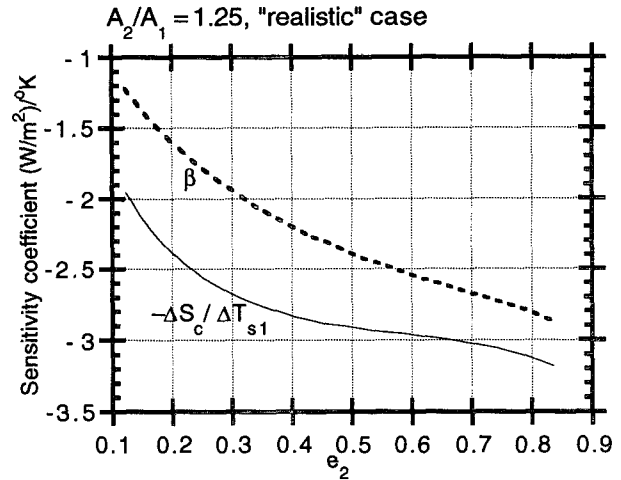

FIG. 12. Sensitivity factors for warm pool sea surface temperature, for $A_2/A_1 = 1.25$ and other parameters as for Fig. 10.

Broccoli and Manabe (1987) report about a 1°C tropical SST reduction due to the 1.2 W m^{-2} radiative change due to reducing CO_2 to 200 ppm in their fixed-cloud simulation; this implies a tropical sensitivity coefficient of -1.2 W m^{-2} K^{-1}. Our model has $\beta \approx -2.5$ W m^{-2} K^{-1} for earthlike climates and thus is roughly half as sensitive as the GCM. The reduced sensitivity of our model is attributable to the fixed cold pool emissivity. The GCM behaves somewhat like a system with fixed relative humidity. Turning attention now to warmer climates, our model gives a 1.6 K tropical warming from the 4 W m^{-2} radiative perturbation caused by doubling CO_2, which is within the middle of the range of GCM predictions and much greater than the 0.5 K warming predicted by Sun and Lindzen's (1993a) model incorporating a stabilizing water vapor feedback.

6. Conclusions

Evaporation does not serve as a thermostat for the tropical climate, even though it makes up a large part of the surface energy budget of tropical oceans. In fact, it is so large that it merely acts as a buffer to keep the sea surface temperature from ever getting much larger than the low-level air temperature. The air temperature itself is determined by wholly different processes. Nor can clouds serve as a primary regulator of the tropical climate, so long as cloud albedo and greenhouse effects cancel in the top-of-atmosphere radiation budget, as they nearly do in the present-day Tropics. When the consequences of the cancellation are drawn out in a physically consistent way, it is found that in equilibrium the clouds have only minor effects on the net column atmospheric heating, the atmospheric circulation, and the warm pool sea surface temperature. Our con-

clusions with respect to clouds are diametrically opposed to those arrived at by Ramanathan and Collins (1991), who argue that clouds have a strong stabilizing effect. To be sure, we have made many simplifying assumptions in our analysis, but the conclusion of Ramanathan and Collins arises from fundamental flaws in their conception of the operation of the tropical circulation system, which are not likely to be resolved in their favor by any amount of additional field study or
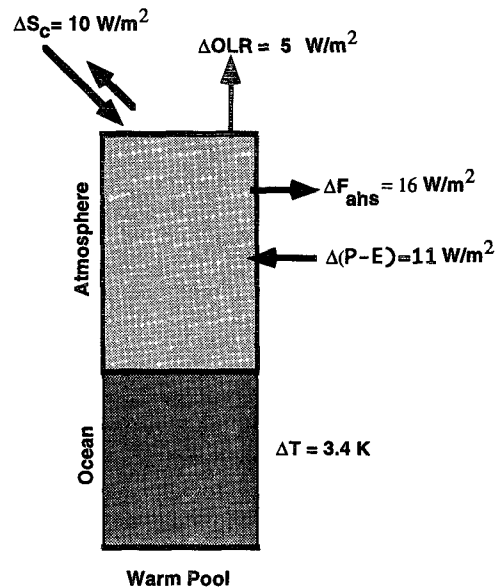

Warm Pool

FIG. 13. Changes in the warm pool heat budget caused by changing S_c by 10 W m^{-2}. Calculation carried out with $e_2 = 0.6$, $A_2/A_1 = 1.25$, and other parameters as for Fig. 10. All fluxes are given as positive quantities, with the arrows indicating whether the flux is a heat export or import.

observational analysis. This does not compromise the value of their presentation of the observations, which provide critical information about cloud optical prop-, erties and their fluctuations.

We are not claiming that clouds can never, under any circumstance, have a palpable effect on the tropical climate. Even if cloud shortwave and longwave forcing cancel at the top of the atmosphere, clouds could perhaps exert an influence through changing the vertical structure of the heating, through changing the relative humidity of the atmosphere, through changing the lapse rate, or through changing the transient response of the atmosphere–ocean system. These effects are subtle, and it is not even clear whether they are stabilizing or destabilizing. Then, too, the cloud longwave and shortwave forcing do not cancel exactly but instead sum up to a small cooling in the course of El Niño fluctuations. Though the residual is small, it is nonetheless comparable to other small forcings driving the climate, such as the radiative perturbation due to doubling of CO_2. We would hardly wish to argue against the need for a better understanding of cloud effects. However, the case that clouds have a distinguished role in controlling tropical sea surface temperature is not compelling. Clouds must take their place among the myriad other influences the climate is heir to. In looking for *big* cloud effects, the question to go after is whether there are altered climates in which either the cloud albedo or the cloud greenhouse effect become dominant.

If one felt compelled to highlight some one influence with the potency to significantly stabilize or destabilize the tropical climate, it would have to be the ability of the atmospheric circulation to create dry air pools in regions of large-scale subsidence. Further, the ability of the dry pools to cool the tropical atmosphere was found to be governed in large measure by low-level horizontal moisture transports, which in turn are tied in with the intricate physics of the tropical boundary layer. It was seen that the warm pool sea surface temperature is very sensitive to the optical thickness of the dry pools and to the relative area of the dry versus convective regions, A_2/A_1. We have not offered a theory for the expected changes of either of these. Determining the dry pool emissivity requires identification and quantification of the moisture sources that offset the subsidence-induced drying. We believe there is an undiscovered stabilization principle at work, which involves adjustment of A_2/A_1. This is suggested by the fact that the configuration of the dry pools changes drastically during El Niño, but the maximum temperature of the atmosphere warms by only 1 K (cf. Figs. 5 and 6). It is worth noting that a modest increase in A_2/A_1 could produce the cool tropical climates suggested by Guilderson et al. (1994) for the last glacial maximum. A modest decrease in A_2/A_1 can warm the Tropics by several degrees, all other things being equal. This would allow the Tropics of the Eocene and other warm climates to export the heat needed to keep the extratropics ice free, without themselves cooling greatly.

There are many directions to pursue in going beyond the elemental model we have discussed. The most important is to replace the unrealistic fixed temperature profile with a real radiative–convective model in each box [perhaps Emanuel's (1991) scheme, which predicts relative humidity]. It would also be desirable to introduce a bit of dynamics so as to allow the atmospheric temperature to differ somewhat between the boxes. Moistening of the dry pool by transient eddy mixing could perhaps be represented as a moisture diffusion between the boxes. With more realistic physics, one could justify coupling the atmosphere into a simple mixed-layer ocean model. Transient adjustment and the seasonal cycle could then be studied. We note also that while the Tropics are mostly ocean, there is also some land near the equator, and at times in the past there was actually quite a lot of land near the equator. Moist rainforests like the Amazon may sustain enough evaporation to act qualitatively like an ocean surface, but desert regions can have a significant effect on the tropical radiation budget. They have high albedo, they can attain high temperatures because of the limited evaporation, and the dry air above them allows effective radiation to space. It is not unreasonable to speculate that subtropical deserts may exert a large influence on fluctuations of the global climate over geological time.

APPENDIX

Glossary of Notation

H	Energy budget of a column of the atmosphere–ocean system
$F_v(\infty)$	Net vertical flux into the top of atmosphere
$F_v(0)$	Net vertical flux out of the bottom of the atmosphere
F_{ah}	Net horizontal atmospheric heat transport (latent + sensible) into the column (per unit cross sectional area of the column)
F_{oh}	Net oceanic heat transport into the column (per unit cross sectional area of the column)
$I^+(\infty)$	Upwelling infrared flux at the top of atmosphere

$I_c^+(\infty)$ Clear-sky upwelling infrared flux at the top of atmosphere

C_l Cloud longwave forcing measured at the top of atmosphere, as defined by ERBE. Positive values indicate a heating.

S Net absorbed solar radiation measured at the top of atmosphere

S_c Clear-sky net absorbed solar radiation

C_s Cloud shortwave forcing measured at the top of atmosphere, as defined by ERBE. Negative values indicate a cooling.

C_s^* Analogous to C_s but measured at the ground instead of at the top of atmosphere

$T(\zeta)$ Vertical profile of the atmospheric temperature

γ Lapse rate of temperature (6.2 K km^{-1})

ζ_T Tropopause height (16 km)

ζ Vertical coordinate, $-h \ln(p/p_{surf})$ (approx. altitude)

p Pressure

p_{surf} Surface pressure

h Scale factor for vertical coordinate (7 km)

α Proportion of the top-of-atmosphere net solar flux that reaches the ground (0.8)

σ Stefan–Boltzman constant

T_s Surface temperature

E Heat flux out of surface due to evaporation (positive values represent cooling of the surface)

F_{sens} Sensible heat flux out of surface

ρ Density of air at the surface

L Latent heat of vaporization of water

u^* Typical boundary-layer wind speed

q^* Typical fluctuation of boundary-layer water vapor mass mixing ratio

C_d Drag coefficient

q_{sat} Saturation mixing ratio

r Relative humidity of air entrained into the atmospheric boundary layer from above

Q_v Energy added to the atmospheric column due to convergence of the vertical flux

F_{aexp} Net heat exported from the Tropics, per unit area of the Tropics (a constant)

Q Net diabatic heating of the atmospheric column due to tropical sources (includes effects of horizontal latent heat redistribution within the Tropics)

P Heat flux due to latent heat release (precipitation) within the atmospheric column

F_{ahs} Horizontal sensible heat transport in the atmosphere (per unit cross-sectional area)

λ Any parameter affecting the climate (e.g., solar constant or CO_2 concentration)

T_r A reference temperature characterizing the atmospheric temperature profile (e.g., tropopause temperature or low-level air temperature)

β Climate sensitivity coefficient (divide into radiative perturbation to get temperature change)

A_1 Area of the warm pool (furnace)

A_2 Area of the cold pool (radiator fin)

M_1 Vertical mass flux in the furnace region (mass per unit area per unit time, upward positive)

M_2 Vertical mass flux in the radiator fin region (mass per unit area per unit time, upward positive)

C_p Specific heat of air at constant pressure

$\delta\theta$ Dry potential difference between the outflow and the inflow of the Hadley–Walker circulation

ζ_{rad2} Effective radiating level of radiator fin atmosphere

$\bar{T}$ Effective radiating temperature $T(\zeta_{rad2})$ of radiator fin atmosphere

e_2 Effective emissivity of radiator fin atmosphere

q_2^* Water mass mixing ratio of the low-level return current from the Hadley–Walker circulation

r_2 Effective relative humidity of the cold pool low-level outflow air

b $= Lq_2^*/(C_p\delta\theta)$. Stability parameter for feedback by moisture import into the warm pool atmosphere.

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