SOME REMARKS ON MECHANISMS FOR THE REGULATION OF TROPICAL SEA SURFACE TEMPERATURE

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1. Introduction

It hardly needs elaboration that the regulation of tropical sea surface temperature is a central problem in climate, with repercussions ranging from the paleoclimatic domain (e.g. climate of the last glacial maximum vs. the warm climates of the Eocene) through the El Niño fluctuations of the present climate, and beyond to projections of climatic impact of anthropogenic CO_2 increase. My object in this brief communication is to convey some of the substance of the spirited and stimulating exchange of ideas on the subject that took place at the NATO ARW to which this volume is devoted. The dialogue has by no means reached closure, and no doubt, the issues I discuss here will continue to be debated for some time to come. It would be safest to view this as my own ideosyncratic view on the subject, which may be wrong in some of its details, but hopefully wrong in an interesting way — or at least in a different way from previous works on the subject.

The central fact of life in the tropics is the destabilizing water vapor feedback, which makes it difficult for a sufficiently moist atmosphere to radiate away tropical levels of insolation locally, without running away to temperatures far in excess of the observed temperature. In its extreme form, the uncompensated feedback can lead to a "runaway greenhouse," anticipated theoretically by Kombayashi (1967) and Ingersoll (1969) and developed further in the intervening work summarized in Renno *et al.* (1994). In the following I will be principally concerned with major "zero order" effects, and especially with the problem of what counteracts the tropical water vapor feedback and brings the sea surface temperature (SST) down to the vicinity of its presently observed range. There are any number of processes (e.g. lapse rate alteration or trade boundary layer adjustments) which could affect the tropical surface temperature by 2-3^{o}K and changes of this magnitude are unquestionably important with regard to climate during an ice age or in a doubled CO_2 world. If one is interested in that level of accuracy, the physics must be treated meticulously and in detail, and the broad-brush approach I adopt in the following will not suffice. To be fair, it is not clear that the current generation of general circulation models suffices for this level of accuracy either.

There are currently three principal mechanisms on the table with regard to tropical SST regulation. They are evaporative feedback (Newell 1979; see also the review given in Waliser and Graham 1993), cloud albedo feedback (the "thermostat hypothesis of Ramanathan and Collins 1991, hereafter RC), and production of dry air in "radiator fins" (Pierrehumbert 1995, with antecedents in Sun and Lindzen 1993). I will argue against the first two proposals, which predict an extraordinarily stable tropical climate; unlike Sherlock Holmes I make no claims as to the validity of the third by default. The radiator fin model still has some notable missing pieces. It succeeds in counteracting the destabilizing water vapor feedback in the warm pool region, but gives little reason to expect tropical climate to be especially resilient.

2. Basic physics governing the temperature of warm oceans

Consider a single column of the atmosphere-ocean system, with atmospheric temperature profile T(z) and ocean surface temperature T_S. The problem of determining the climate can be divided into two steps: (**PROBLEM A**) the determination of T_S *given* T(z), and (**PROBLEM B**) the determination of the T(z) that satisfies the top-of-atmosphere radiation balance and the constraints of radiative-convective dynamics. Problem B is not as underdetermined as it looks, because the form of T(z) is highly constrained. In the stratosphere, it is uniquely determined by radiative equilibrium and the atmospheric composition. In the troposphere, the situation is somewhat more complicated, since the profile is set by the mixing due to moist convection. Nonetheless, observations and sophisticated models (Sun and Lindzen 1993; Emanuel 1991) indicate that the tropics stays sufficiently close to the saturated moist adiabat for that to constitute a satisfactory "zero order" description. The importance of this is that T(z) is (approximately) constrained to a 1-parameter family of curves, so that specifying the temperature at any one level determines the atmospheric temperature at all levels.

For a warm, moist atmosphere such as prevails in the tropics, changing T_S while keeping T(z) fixed yields little change in the infrared radiation escaping the top of the atmosphere (the "OLR"). This is because such atmospheres are optically thick in the infra-red. In consequence, Problem A decouples from Problem B. One can first determine T(z), then find T_s *ex post facto* by consideration of the surface energy budget.

A major complication to this picture is the necessity of specifying a water vapor and cloud water profile when solving the radiative transfer problem to determine the OLR. These profiles typically depend on many things besides T(z) itself. The classical prescription for water vapor is to fix relative humidity, but other approaches are emerging (Sun and Lindzen 1993, Emanuel 1991). Representation of cloud effects in radiative-convective models is at an even earlier and more exploratory stage of development.

Whatever prescription is used, the aim is to identify a reference temperature T_{ref} characteristic of the atmosphere, determine the OLR as a function of T_{ref}, and then to determine T_{ref} by balancing OLR against the absorbed solar radiation. In its simplest incarnation, one adopts a constant temperature lapse rate Λ, and uses the radiation physics to determine an effective radiating level z_{rad}. The state of the climate is then determined by $S = \sigma T(z_{rad})^4$, where S is the absorbed solar radiation. The low level air temperature is then $T(0) = T(z_{rad}) + \Lambda z_{rad}$. In Section 3, I will argue that evaporation makes the surface budget "stiff" over moist surfaces, so that T_s cannot deviate much from T(0). It thus emerges that the top-of-atmosphere radiation budget together with the lapse rate are the main players in determining the surface temperature.

The primacy of the top-of-atmosphere considerations is important to keep in mind when assessing cloud impacts, because — in the present climate — the cloud albedo and cloud greenhouse effects very nearly cancel each other at top-of-atmosphere in the tropics (RC). A complication is that the atmosphere is not in local radiative equilibrium in the tropics, so that one must allow for lateral heat transports into and out of the atmosphere-ocean column. In Section 3, I will summarize some arguments showing that, in equilibrium, these transports do not substantially alter the naive expectation that cloud effects approximately drop out of the determination of atmospheric temperature.

3. Some illustrative models

First let's consider the sensitivity of SST to removal of clouds and to changes in the oceanic heat flux *assuming that the atmospheric temperature remains fixed during these changes*. This is a rather unrealistic thought experiment, in that climate change will generally involve changes in the atmospheric temperature T(z), rather than just changes in surface temperature. However, there is much to be learned about the interplay of clouds and evaporation from this restricted problem; later, I will bring atmospheric temperature changes back into the picture.

The temperature fluctuations of a mixed layer ocean with mixed layer depth h are governed by

$$\frac{dT_s}{dt} = F_{surf} / (\rho_w C_{pw} h) \qquad (3.1)$$

where T_s is the surface temperature, ρ_w is the density of sea water, and C_{pw} is the specific heat of the sea water. The surface layer heating is given by

$$F_{surf} = \alpha S_c + \gamma C_s + \{I^-(0) - \sigma T_s^4\} - E - F_{sens} + F_o \qquad (3.2)$$

S_c and C_s are the clear-sky absorbed solar flux and the cloud shortwave forcing measured at the top of the atmosphere, as defined by ERBE (cf Ramanathan *et al* 1989). Further, α is a clear sky shortwave absorption coefficient representing the portion of the clear sky top-of-atmosphere solar flux which reaches the ground without being absorbed by the intervening atmosphere, and γ is a coefficient measuring the effects of cloud shortwave absorption on surface insolation. $I^-(0)$ is the downwelling infrared flux impinging on the surface, E is the evaporative heat flux, F_{sens} is the sensible heat flux, and F_o is the heat added by oceanic transports into the surface layer (negative values signify cooling by importation of cold water, from the sides or from below the mixed layer).

Let's represent evaporation by the bulk aerodynamic formula (Peixoto and Oort 1992)

$$E = \rho L\, C_d\, u^*\, q^* \qquad (3.3)$$

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. q^* can be estimated using

$$q^* = q_{sat}(T_s) - r\, q_{sat}(T(0)) \qquad (3.4)$$

in which q_{sat} is the saturation mixing ratio and r is the typical relative humidity of air entrained into the boundary layer. Following RC, I empirically represent the cloud shortwave forcing as

$$C_s = -a\,(T_s - T_{crit}) \quad \text{for} \quad T_s > T_{crit} \tag{3.5}$$

with a = 20 $W/m^2\,{}^oK^{-1}$. C_s is set to zero below the T_{crit} = 300^{o}K convective threshold, and saturates at the temperature where all the solar radiation is blocked. To simplify the discussion, I shall ignore F_{sens}.

I will now discuss some transient adjustment problems carried out with the above system, with the atmospheric temperature held fixed. The fixed-atmosphere assumption is implemented by specifying T(0) constant at 300^{o}K and $I^-(0)$ constant at 393W/m^2, which is roughly the back radiation that corresponds to T(0) with a relative humidity at 80%. The other fixed parameters are: h = 40m, S_c = 370 W/m^2, α = .8, u* = 4 m/s, r = 80%, and C_d = .0015. Based on the growing evidence for anomalous cloud shortwave absorption, we set γ=1.5 (Ramanathan, *et al* 1995), though this value should at present be regarded as somewhat speculative. With these parameters, the SST equilibrates at 302.1^{o}K if F_o is set to -30 W/m^2. The equilibrium evaporation for this case is 114 W/m^2, which is entirely consistent with the observed mean value in the Pacific warm pool (Ramanathan *et al* 1995) if one allows 10 W/m^2 for the neglected sensible heat flux.

In the first adjustment experiment, one begins with the equilibrated system and instantaneously removes the clouds[1] by setting a=0. The adjustment of T_s and the evaporation is shown in Figure 1. Note that even complete removal of the cooling effect of clouds results in only a 2.3^{o}K rise in the SST. This is because the excess insolation caused by removal of the clouds is accomodated mostly by a 60 W/m^2 increase in the evaporation, which takes about 100 days to come into being. The rise in evaporation does not require a great increase in SST, because evaporation is such a steeply increasing function of the difference between the SST and the overlying air temperature.

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[1] Properly speaking, one ought to reduce $I^-(0)$ somewhat when removing the clouds, but because of the optical thickness of a warm moist atmosphere, this effect is rather inconsequential in comparison with the effect of clouds on surface insolation. Our calculation slightly exaggerates the warming effect of removing clouds, and therefore, if anything, overestimates the role of clouds.

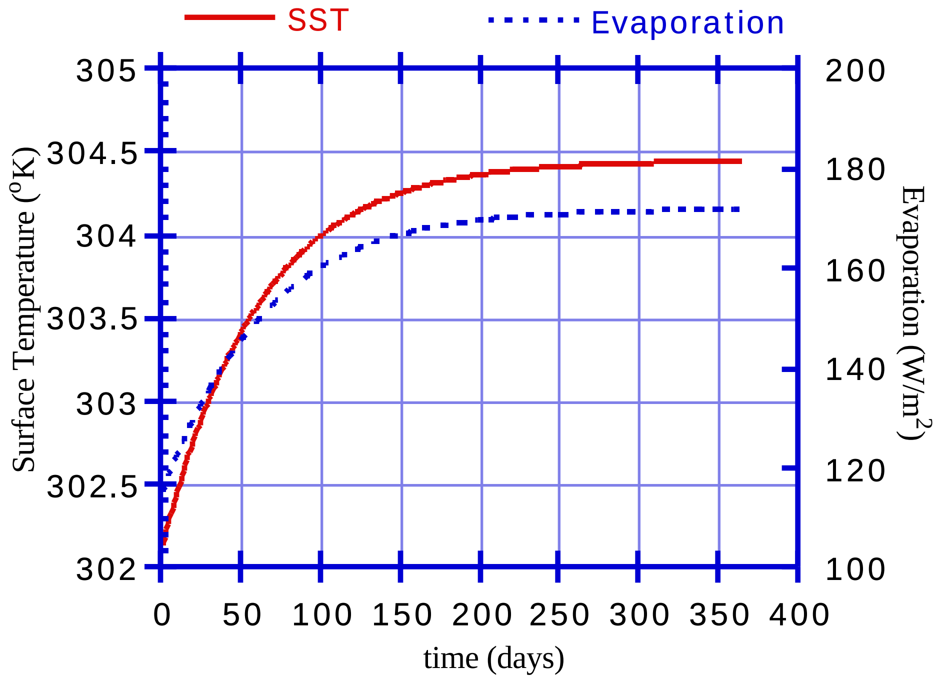

Figure 1. Adjustment of sea surface temperature and evaporation, following elimination of clouds. Calculation carried out with atmospheric temperature held fixed.

Clearly a 2.3°K change is not insignificant. In regions of light winds, clouds undoubtedly help hold down the air-sea temperature difference. My point is that the large effect of cloud shading on surface insolation can give an exaggerated notion of the sensitivity to clouds, if the buffering influence of evaporation is not taken into account. Modest increases in the wind speed or drag coefficient sharply reduce the impact of clouds. If we increase C_d u* from .006 to 0.01, the SST with clouds is 301.12, and without clouds is 302.02 — yielding a cloud effect of only .9°K.

To be sure, when clouds are eliminated, the evaporation rises to values far in excess of the long-term mean values that currently prevail over the warmest Pacific waters. This is no problem, because I am not claiming that there are no clouds in the Tropics — only that the air sea temperature difference wouldn't increase terribly much if clouds were eliminated. Moreover, the flatness of scatter plots of observed evaporation vs. SST (Zhang and McPhaden, 1995) poses no contradiction to the kind of evaporation model used in computing Figure 1. The analysis of Zhang and McPhaden implies only a *correlation* between high temperatures and the low windspeeds that limit evaporation. It does not imply that increasing the SST by any means and at any place will inevitably reduce winds. Given that the tropical winds are determined by aspects of the general circulation (like the Hadley cell) spanning the entire tropics, there are good reasons to doubt a local, causal connection. It seems likely that low winds cause high SST, rather than the other way around.

As an exercise in sensitivity, I next compare the adjustment of SST upon elimination of the oceanic cooling for the cloud and the no-cloud case. In both cases the SST is initialized with the equilibrium temperature for F_o = -30 W/m^2. Results are shown in Figure 2.

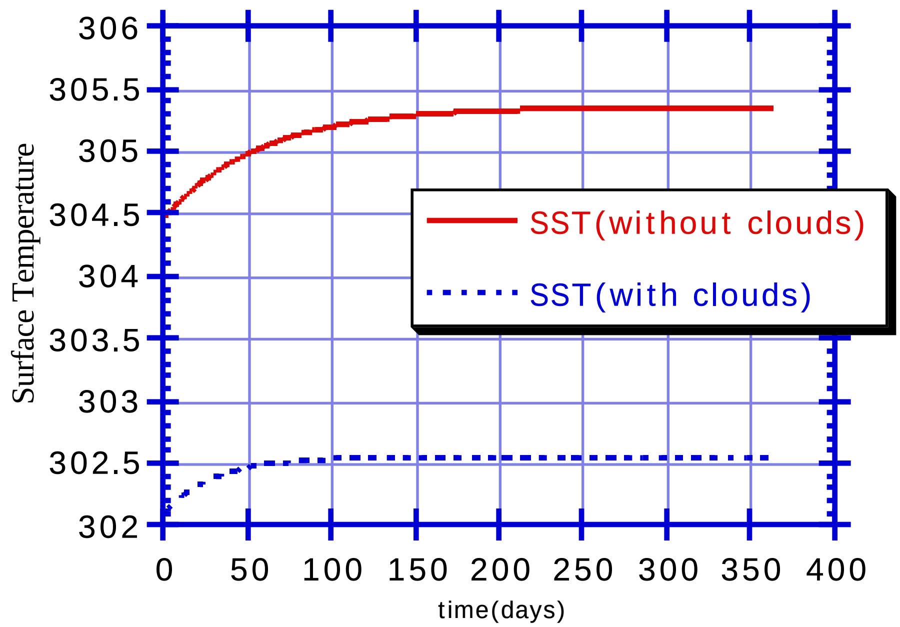

Figure 2. Adjustment of sea surface temperature after removal of the 30 W/m^2 oceanic cooling. Results are shown with and without clouds.

Because of the influence of evaporation, the difference in sensitivity between the two cases amounts to just .45^{o}K.

From the preceding one concludes that evaporation has a powerful buffering effect on the surface energy budget, making the air-sea temperature difference resistant to large changes in the terms in the budget. Lindzen, Hou and Farrel (1982) also noted this effect. My argument is in essence the same given by Newell (1979), save that Newell overstated the implications of the result because he failed to take into account the fact that climate perturbations generally change T(0). *I am not arguing that evaporation exerts the primary control on the SST itself.* It keeps SST from much exceeding T(0), but does nothing to keep the atmospheric temperature from running away to large values. To determine T(0) itself, one must look to the top-of-atmosphere energy budget.

It might be thought that the preceding calculation is irrelevant to the debate on the thermostat hypothesis, because of the assumption of fixed atmospheric temperature. In fact it is very generous to the operation of a cloud thermostat , since the cloud longwave forcing C_l is prevented from warming the atmosphere, and from warming the surface. In reality, the atmosphere will warm "some" in response to C_l. How much? The answer to this question requires an estimate of the lateral heat flux in the atmosphere, which can carry away some of the excess heating of the atmospheric column and allow it to equilibrate at a lower temperature than would be the case for an isolated column.

The atmospheric heating which forces the atmospheric circulation can be written

$$Q_{atm} = \{\sigma T_s^4 - I^-(0)\} + (1-\alpha)S_c + (1-\gamma)C_s + Q_{L1} + C_l - I_c^+(\infty) \quad (3.6)$$

where $I_c^+(\infty)$ is the clear-sky OLR extrapolated to the whole scene as defined by ERBE, Q_{L1} is the latent heating of the column *that would result from local evaporation from the ocean below the column*, and the rest of the symbols have the same meaning as before. Note that (3.6) does not contain terms for latent heating due to redistributed moisture, nor for adiabatic cooling due to mean ascent; these are considered as part of the lateral atmospheric heat transport, since they are tied to the circulation driven by the heating. The atmospheric equilibrium temperature is determined by balancing Q_{atm} against the atmospheric horizontal heat transport F_{ah}. The budget (3.6) involves C_l but does not appear to involve all of C_s , which appears explicitly only in the solar absorption term. However, the rest of C_s can work its way back into (3.6) in a sneaky way. The change in surface insolation implied by C_s results in a change in the terms σT_s^4 and Q_{L1} (which equals E in a convecting region) as the surface budget attempts to come into equilibrium. For example, the dominant balance in the surface budget is between insolation and evaporation, so that reductions in insolation caused by increasing C_s eventually lead to strong reductions in E and Q_{L1} (Washington and Meehl 1993, Boer 1993). It was shown in Pierrehumbert 1995 that regardless of the detailed form of the terms in the surface energy budget, the cloud effects show up in Q_{atm} only in the combination C_s+C_l, once the surface budget comes into equilibrium.

The preceding considerations apply even if the atmospheric transports are computed using a full GCM, but further insight can be gained using the observation that atmospheric heat transports in the tropics are effective enough to keep the atmospheric temperature nearly horizontally homogeneous. This means that T(z) is determined by an average of Q_{atm} over the whole tropics rather than by its local behavior (except for behavior in a shallow boundary layer, which can tolerate more geographical variation because of the weak pressure gradients induced by shallow temperature anomalies.) Thus, T(z) and the moisture profile need to be adusted so as to satisfy

$$\langle Q_{atm} \rangle = 0 \qquad (3.7)$$

apart from the effects of atmospheric heat export to midlatitudes. The implied lateral heat flux within the tropics is $F_{ah} = \langle Q_{atm} \rangle - Q_{atm}$, and thus depends on all the heat sources enumerated in (3.6), and also on the behavior of the terms over the whole tropics, not just in the warm pool region. In contrast RC, who also invoke the uniform temperature approximation, argue that $F_{ah} = -(1\text{-}f)C_l$, for some small factor f; this estimate is sensitive only to the local cloud longwave heating of the atmosphere.

In order to illustrate some of the consequences of these ideas, I will now extend the earlier SST adjustment model to allow for changes in the atmospheric temperature. In this model I will make some simplifying assumptions which are not quantitatively accurate, but which enable us to focus on the main players while retaining physical consistency. First, I assume that the atmosphere is transparent in the visible (α=1, γ=1) and that the net IR cooling of the surface $\{\sigma T_s^4 - I^-(0)\}$ is negligable; corresponding terms are also dropped from the surface budget, to preserve consistency. To evaluate (3.7) the system is expanded about a presumed equilibrium with $T(0) = 300^oK$, so that the clear sky OLR term is approximated by

$$\langle I_c^+(\infty) \rangle = OLR_o + \beta_1 (T(0) - 300^oK) \qquad (3.8a)$$

where β_1 is some constant characteristic of the moisture distribution over the whole tropics. The "warm pool" region is assumed to occupy a fractional area A<1 of the tropics, and the changes in Q_{L1} relative to the base state are assumed to vanish outside this region. Similarly, the cloud effects are limited to the warm pool region. With these assumptions, (3.6,3.7) can be expanded around equilibrium to yield

$$0 = \langle Q_{atm} \rangle = - \beta_1 (T(0) - 300^oK) + A\{E-E_o + C_l-C_{lo}\} \qquad (3.8b)$$

where E_o and C_{lo} are the equilibrium evaporation and cloud longwave forcing corresponding to T(0) = 300^{o}K. The evaporation and cloud forcing have been assumed horizontally uniform over the warm pool region. E and C_l are determined as in the previous SST adjustment model, with the further assumption that $C_s(T_s)+C_l(T_s) = 0$. The surface temperature is determined as before by integrating (3.1) with the appropriate terms dropped from (3.2). With (3.8b), the changes in T(0) needed to keep the atmosphere in equilibrium in the face of changing E and C_l can be found; effects due to the finite adjustment time of the atmosphere are ignored, though some calculations (Emanuel 1991) suggest that the water vapor content of the atmosphere can take quite a long time to come into equilibrium. The calculations below were carried out with β_1 = 2 W/(^{o}Km^2). More comprehensive models of the equilibrium state without so many limiting assumptions, and including an explicit computation of β_1 in terms of a radiative model of the dry, subsiding regions of the tropics, are given in Pierrehumbert 1995.

In the first experiment, I take a system which is initially in equilibrium with F_{oh} = -30 W/m^2, and then instantaneously eliminate the clouds. The subsequent adjustment of the atmospheric temperature is shown below, for two different choices of the warm-pool area.

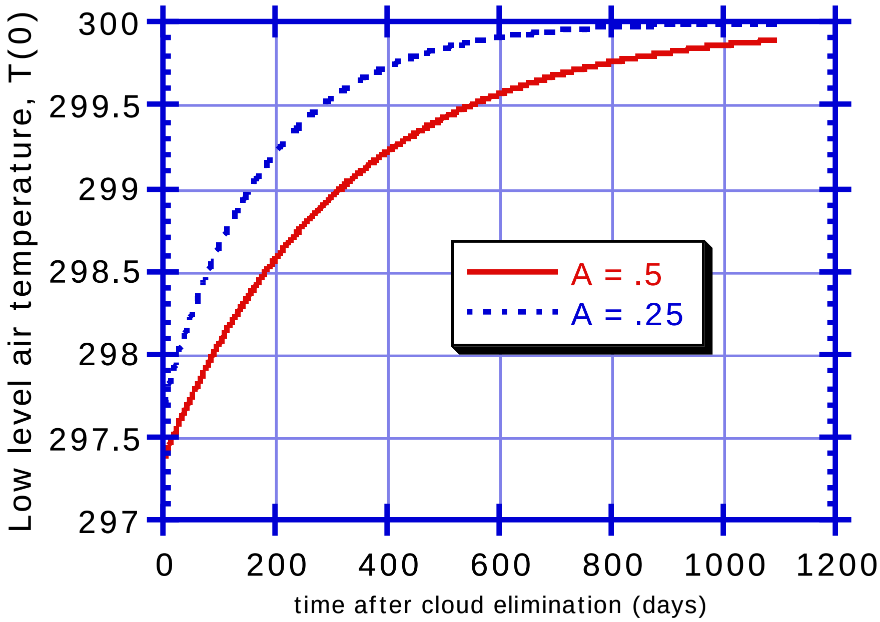

Figure 3. Adjustment of atmospheric temperature following instantaneous elimination of clouds.

Initially, the atmospheric temperature *drops* several degrees. This is because the cloud longwave heating C_l has been removed from Q_{atm}, but the evaporation has not yet had time to adjust to bring the surface budget back into equilibrium. Because a heating term has been removed, the atmospheric temperature must go down so as to reduce the OLR and bring the atmospheric budget into balance. As time goes on, however, the surface comes back into equilibrium, the evaporation rises to compensate the elimination of C_s from the surface budget, and the resulting change exactly compensates for the elimination of C_l — supposing, as always, that $C_s+C_l = 0$. In the end, the atmosphere returns to the same 300^{o}K equilibrium temperature it had in the presence of clouds, though it requires a year or so for the recovery to take place.

In the next thought experiment, I consider the adjustment of the atmosphere to a new equilibrium following elimination of the 30 W/m^2 oceanic cooling in the warm pool region alone. The calculation is performed with and without clouds, but in each case the surface temperature is initialized in equilibrium with the 30 W/m^2 cooling prior to its elimination. Hence the atmosphere is initially in equilibrium at its unperturbed temperature of 300°K. Results are shown in the following graph.

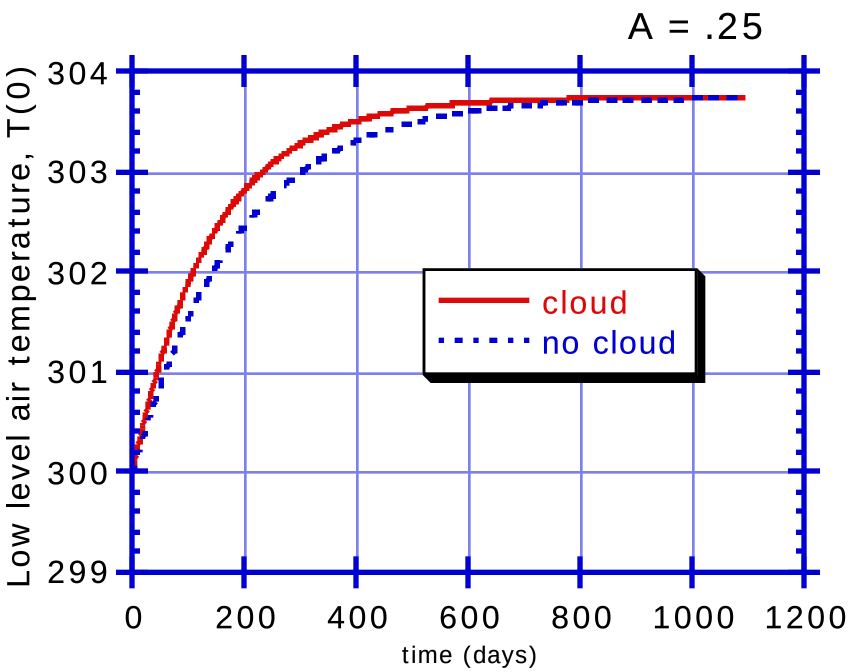

Figure 4. Low level atmospheric temperature adjustment following elimination of 30 W/m^2 oceanic cooling., with and without cloud feedback. Both calculations were initialized with the equilibrium corresponding to the 30 W/m^2 cooling. The convective region was assumed to occupy 25% of the area of the tropics.

Clouds affect the rate of approach to the new equilibrium, but eventually the surface budget comes back into balance so that clouds have no effect on the atmospheric temperature that is ultimately attained. In both cases, the atmospheric warming caused by elimination of F_o is 3.75 °K, which is fixed by averaging over the whole tropics the 30 W/m^2 change in warm-pool oceanic cooling (yielding 7.5 W/m^2 for A = .25), and dividing by the clear-sky OLR coefficient β_1.

In equilibrium, clouds thus have little impact on the atmospheric temperature, insofar as $C_s+C_l = 0$. If clouds are instantaneously removed from a system in equilibrium, elimination of clouds actually *cools* the atmosphere, during the time the surface budget is out of equilbrium. Conversely, sudden introduction of clouds to a cloud-free atmosphere would temporarily *warm* the atmosphere, making it harder for any cloud thermostat to prevail. If lateral heat transports are effective enough, the atmosphere may not warm *much*, but if the atmospheric warming becomes negligable we are back in the domain of the fixed-atmosphere calculation.

The sensitivity of the warm pool sea surface temperature to perturbations can be summed up in the sensitivity coefficient[1] β, defined so that

$$\Delta T_s = \frac{\Delta F}{\beta} \tag{3.9a}$$

where ΔF is the imposed alteration of the energy flux into the atmosphere-ocean column and ΔT_s is the shift of warm pool SST in response. For example ΔF could be the roughly 4 W/m^2 radiative forcing due to doubling CO_2, or the roughly 25 W/m^2 shift in oceanic heat flux that might be expected in the course of El Niño fluctuations. In Pierrehumbert (1995) it is argued by systematic application of surface equilibrium, the uniform-temperature approximation, and the C_s+C_l cancellation that

$$\beta = \frac{\partial \langle I_c^+(\infty)\rangle}{\partial T} \tag{3.9b}$$

where T represents the atmospheric temperature, $I_c^+(\infty)$ is the *clear sky* outgoing longwave radiation as defined in ERBE, and angle brackets denote a horizontal average over the tropics. This stability estimate ignores the possibility of stabilizing or destabilizing feedbacks due to systematic shifts of heat exports from the tropics to the extratropics, and also neglects the modest effects of clouds on the sensitivity of the air-sea temperature difference. The two-box model of the tropics discussed in Pierrehumbert (1995) yields $\beta \approx 2\ W/({}^oK m^2)$. Owing principally to their different assumption about lateral atmospheric heat transport, RC's estimate of sensitivity is dominated by the $\partial C_s/\partial T_s$ term, and yields sensitivity coefficients upwards of 15 $W/({}^oK m^2)$, suggesting an extraordinarily stable tropical climate. Clearly, one of these estimates is seriously in error.

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[1] There is a difference in sign between the definition of β employed in RC and that used in Pierrehumbert (1994). Here, we adopt RC's sign convention.

4. Prospects for clouds as a thermostat?

An issue of general importance to the effect of clouds on climate is the dependence of cloudiness on atmospheric temperature. To be sure, ERBE results (see esp. RC) show a dramatic increase in tropical cloud forcing as the SST exceeds a trigger temperature of about 300^oK. However, this critical temperature does not represent an eternal thermodynamic constant for the planet, but rather the temperature necessary for the onset of deep convection. Deep convection is possible when the moist static energy of the low level air imparts sufficient buoyancy for air parcels to reach the upper troposphere, and so the critical temperature would increase if the atmosphere became warmer. With a warmer atmosphere a cloud thermostat, if any, would not even begin to kick in until higher temperatures, and so it is hard to see how clouds could exert a fundamental control on the atmospheric temperature. A nice illustration of this effect is provided by the GCM studies of Washington and Meehl (1993), carried out with a mixed layer ocean. Owing to the neglect of oceanic heat transports, their model equilibrates at a warmer tropical temperature than is the case for the real atmosphere; in accord with this, the threshold for onset of deep convection (as measured by the sharp rise of the atmospheric greenhouse effect) is 304^oK rather than 300^oK. Thus, the use of a fixed cloud-SST relation in my models of atmospheric adjustment discussed in the previous section is not strictly consistent, and may exaggerate the effects of clouds on the transient behavior.

Do we expect a generally warmer climate to be more or less cloudy than the present one? Given sufficient moisture supply one could argue (though not necessarily correctly) that a warmer atmosphere would be generally moister, owing to the Clausius-Clapeyron relation. There are no such simple limits for condensed cloud water, however. Cloudiness depends on the production of saturated air, not on the water vapor content of air. The condensate burden is limited by settling, evaporation into unsaturated air, and coalescence/precipitation. None of these effects is primarily controlled by temperature. In fact, if warmer air were generally more cloudy, we would expect the warm low level air to be densely swathed in clouds. Cirrus clouds would be practically nonexistent.

Dynamics as well as microphysics plays a crucial role in determining the the source of high altitude clouds. Dynamical transports keep T(z) relatively uniform over the whole tropics, which means that it cannot warm locally to accomodate local anomalously warm SST's. This process keeps T(z) cool enough to permit vigorous, deep convection in localized convective regions, at the expense of suppressing convection elsewhere. This modulation is supported further by the direct effect of the ascending and subsiding branches of the Hadley/Walker circulation, which powerfully enhance or suppress convection in the conditionally unstable atmosphere. Very deep convection provides the water vapor injection needed to create a dense upper tropospheric cirrus shield. The cloud patterns would be very different in a world with more uniform tropical SST's, which would entail more ubiquitous though less intense convection.

I have been deliberately vague in using the term "cloudiness" above. In reality, it is not just cloudiness, but the imbalance between cloud longwave and shortwave forcings that count. What becomes of the the approximate cancellation between C_s and C_l when we go to a generally warmer (2xCO_2, Eocene or Cretaceous) or cooler (ice age) atmosphere? This stands as the most pressing unresolved question concerning cloud effects on climate.

5. Conclusions

According to my picture, the cloud thermostat hypothesis, as stated in RC has the following three shortcomings:

- Viewed from the standpoint of the surface energy budget, it doesn't take into account the large evaporation term, which can change and "buffer" changes in cloud shortwave forcing.

- From the standpoint of the atmosphere-ocean column budget, RC's estimate of lateral atmospheric heat flux errs in not properly taking into account all heatings which drive the atmospheric circulation, and in not taking into account the importance of heating *contrasts* across the tropics. If there is not enough cooling somewhere in the tropics, the excess heating in the warm pool region cannot be gotten rid of, no matter how effective the lateral heat transport, and the tropics as a whole would have to warm significantly. The chief problem, though, centers on the behavior of latent heating in the warm pool atmosphere. In studying this issue through observations or GCM simulations, one must take care to distinguish transient effects from equilibrium behavior, since the latent heating adjustment will not be instantaneous.

- The critical surface temperature above which deep convection and tropical cloudiness increase is not a universal constant, but instead is expected on elementary physical grounds to increase or decrease in step with the overall atmospheric temperature of the tropics. Therefore, it is a fallacy to extrapolate to perturbed climates using a fixed critical temperature.

I am 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 perhaps 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. 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. I would hardly wish to argue against the need for a better understanding of cloud effects.

Based on the above considerations, I would like to propose a "modified thermostat hypothesis." It is a considerably weaker and less dramatic assertion that the original one contained in RC, but it is what I think is actually supported by the CEPEX and ERBE data. The **Modified Thermostat Hypothesis** would read something like the following: "Insofar as surface winds tend to be weak over the warmest parts of the tropical ocean, evaporation alone cannot account for the tight coupling between SST and the overlying air temperature. In these regions, enhanced cloudiness plays a critical role in limiting the air-sea temperature difference." Without clouds, the air sea temperature difference in the warm pool region might be 2^oK larger than the observed value prevailing in the current climate.

Evaporation and clouds may jointly control the air-sea temperature *difference*, but this does not provide a fundamental limit to the tropical surface temperature itself, which is determined primarily by top-of-atmosphere considerations. While the true extent to which tropical SST has varied over geological time is still an unsettled issue, there are indications that the variations have been quite limited, particularly on the warm side. If these indications hold up, what could account for the apparent stability? I think the most likely candidate is the propensity of the large scale tropical circulation to create extensive pools of very dry air in undisturbed subsiding regions. These "radiator fins" are coupled dynamically to the rest of the tropics, and allow excess insolation to escape to space. Elsewhere (Pierrehumbert 1995) I have attempted to study their impact quantitatively, but much remains to be done before this problem can be laid to rest.

In my opinion, the original conception of CEPEX as a test of the thermostat hypothesis of RC has outlived its usefulness, in light of what has been learned by further thinking about the problem of SST regulation. This doesn't in any way compromise the importance of the observations taken during CEPEX. In fact, the outstanding success of CEPEX stands as a vindication of the concept of hypothesis-driven field experiments. The CEPEX data has already proved its worth in addressing questions related to cirrus lifetime and optical properties, and the factors governing the moisture distribution of the tropics. By any reckoning, such questions must figure prominently in any treatment of the effect of clouds in an altered climate.

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Notes for revision

From my email exchange with Coakley

Fa is the atmospheric lateral heat flux, if we're agreed on the notation. One has to distinguish between the forcing and the response. The unbalanced atmospheric heating Qa is the FORCING for the atmospheric transport Fa. If Qa = 0 everywhere, we have local radiative-convective equilibrium of the atmosphere, with no need for any circulation. Including Fa in the heating Qa would bury the very distinction that makes it possible to estimate the atmospheric lateral transports. Qa is exactly the heating one would have to put into a numerical model of the tropical atmosphere in order to explicitly drive the circulation.

There are some slightly tricky parts to this argument, relating to how you account for latent heat from imported water vapor, but that's explained more clearly in the new version.

To be sure, if one started with the atmosphere in the balance it is in today, and suddenly removed the clouds keeping everything else fixed, the surface would be 'way out of equilibrium. That thought experiment has some relevance to the transient adjustment problem, but I don't see what it has to do with the problem of the new equilibrium the system would find without clouds (which, I argue, is not too different from the present equilibrium). The system would adjust by the surface warming a bit, and just how much that "bit" is depends on the behavior of evaporation with temperature. Note that it would be a gross error to take the observed evap(SST) curve from, say Zhang's paper, and assume that this is the relevant curve for the no-cloud thought experiment. Zhang's curve already reflects the fact that the real atmosphere has clouds, which make less evaporation necessary. To do the no-cloud experiment,

you need a model of evaporation. My drag-law model yields
a prediction of rather small cloud effects EVEN IF THE ATMOSPHERIC TEMPERATURE IS HELD FIXED. I later found that Lindzen, Hou and Farrell got a similar result in a more realistic radiative-convective model, so it's not just my simplifications that are doing the trick.

The second part of the argument deals with how much the atmosphere might warm up, which would allow the SST to warm further. That is inextricable from the heat flux issue. My "no cloud" effect on Qa is in fact rather favorable to a thermostat, in the sense that it allows Cl to NOT WARM THE ATMOSPHERE AT ALL. The rest of my argument essentially deals with the issue of what actually determines the tropical atmospheric temperature, and I conclude that it is actually the humidity distribution that is the main player.

Here's another thought experiment. Start with the system in equilibrium,
with clouds as observed. Now, instantaneously wipe away the clouds. The system is now not in equilibrium. What happens? First, because we no longer have Cl, the OLR increases and the atmosphere experiences a net cooling. The heating Qa was formerly in balance with the circulation Fa, but now the circulation must change to accomodate the additional localized cooling. The atmosphere will cool SOME, depending on the effectiveness of the horizontal heat transports. However, at the same time, we must look at what's going on at the surface. At the initial instant, since we've changed the clouds without changing SST or atmospheric temperature, the evaporation is unchanged. However, at the same time the atmosphere starts to cool, the surface will start to warm because we've eliminated Cs. With any reasonable evaporation model, this situation will rapidly cause the evaporation to increase;

when the system comes into a new equilibrium, the increased evaporation

leads to increased latent heating, which offsets the cooling caused by eliminating Cl. In the new equilibrium, the surface temperature is "a bit" higher, as I described above. Note that actually, I've oversimplified by talking only about absorbed solar and evaporation here, but in fact the argument goes through in the same way if you put in all the heat fluxes. It also doesn't matter if there is solar absorption in the atmosphere, since that just means that the heat reaches the atmosphere directly rather than through the intermediary of evaporation. Solar absorption does change the details of the transient adjustment process, though.

Notes on clouds in a warm atmosphere

How would clouds behave in a generally warmer atmosphere? A warmer atmosphere can hold more moisture, but it won't necessarily hold more liquid water. If cloudiness increased with water vapor content, then clouds would be mostly in the tropical boundary layer, rather than in the cold cirrus layer. Of course, it is the proportion of saturated air that determines cloudiness. Deep convection is good at producing clouds because it pumps a great deal of water into the cold tropopause region. But what determines the liquid water content of a level of the atmosphere? (Balance between moisture flux, and sink of condensate due to precipitation (via coalescence or direct settling), and evaporation.). Is there an upper bound on the liquid/ice water content of the atmosphere? How does it depend on temperature?

What happens to clouds in a 1D radiative convective model? What takes the place of the "trigger temperature" scaling?