Dynamic ocean-atmosphere coupling: a thermostat for the tropics

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Abstract

The ocean currents connecting the western tropical Pacific ocean with the eastern tropical Pacific ocean are driven by surface winds. The surface winds are in turn driven by the sea surface temperature differences between these two regions. This dynamic coupling between the atmosphere and the ocean may limit the sea surface temperature in the tropical Pacific ocean to below 305 K even in the absence of cloud feedbacks.

Records for the past climates and observations of the interannual fluctuations about the present climate suggest that the maximum tropical sea surface temperature (SST) is somehow limited to 305 K (1, 2). On the basis of data from the Earth Radiation Budget Experiment, Ramanathan and Collins (3) hypothesized that the tropical Pacific SST is mainly regulated by a negative feedback from cirrus clouds, a proposal they referred to as the "thermostat hypothesis". This hypothesis, however, is controversial (4-8).

In this article, we propose an alternative thermostat for the SST in the tropical Pacific ocean: the dynamic coupling between the atmosphere and the ocean. The ocean currents that connect the western Pacific ocean with the eastern Pacific ocean are driven by surface winds. Surface winds are in turn driven by the SST difference between these two regions (9). This coupling plays a central role in the ENSO phenomena (2, 10), but its importance for the mean tropical climate has been less clear.

To illustrate the mechanism by which the dynamic ocean-atmosphere coupling regulates tropical SST, we consider a three box model for the tropical Pacific ocean coupled with a simple atmosphere (Fig. 1). The surface Pacific ocean over the equatorial region is represented by two boxes with temperature T_1 and T_2. The two boxes are assumed to have the same volume. The subsurface ocean is represented by another box with temperature T_c. Using T_e to represent the radiative-convective equilibrium temperature, a temperature that the surface ocean would attain in the absence of the east-west ocean current, and c the reciprocal of the time scale for the radiative-convective processes, we may write the heat budget of the two boxes for the surface ocean as follows:

$$\frac{dT_1}{dt} = c(T_e - T_1) + q(T_2 - T_1) \tag{1}$$

$$\frac{dT_2}{dt} = c(T_e - T_2) + q(T_c - T_2) \tag{2}$$

with q given by

$$q = \alpha(T_1 - T_2) \tag{3}$$

The first term on the right hand of (1) and (2) is the heat exchange with the atmosphere, and the second term the advection of heat by the ocean current. In deriving Eq. (3), we have assumed that the strength of ocean current is proportional to the strength of the surface wind, and that the strength of the surface wind is proportional to the temperature difference between the east and the west (9). α is a constant that is related to eddy damping time scale in the atmospheric boundary layer and in the mixed layer of the ocean. Because the simple coupled system does not contain any physics that distinguishs the east from the west, we have also assumed $q \geq 0$. To simplify the analysis, T_c is assumed to have a fixed value. It has been shown that the heat deposited from the surface ocean to the subsurface ocean over the equatorial region can be effectively removed to the subtropical ocean through the meridional branch of the wind-driven circulations (12-14). The merdional branch consists of the equatorial upwelling, the poleward Ekman drift, and the subtropical subduction. When the subtropical ocean is included in the box for the subsurface ocean, the meridional circulaion is implicitly taken into account in this two dimensional representation.

The behavior of the simple coupled system constituted by Eqs. (1)-(3) is completely determined by a single nondimensional parameter $\alpha^* = \frac{\alpha}{c}(T_e - T_c)$. For a fixed sensitivity of ocean currents to east-west SST difference, α^* directly measures the capacity of the atmospheric feedbacks in trapping heat. T_e and c are given by the following equations,

$$T_e \simeq T_0 + H_0[\frac{\partial E}{\partial T} - (\frac{\partial G_a}{\partial T} + \frac{\partial C_l}{\partial T} + \frac{\partial C_s}{\partial T})]^{-1} \tag{4}$$

$$c = \frac{1}{C_p \rho h}[\frac{\partial E}{\partial T} - (\frac{\partial G_a}{\partial T} + \frac{\partial C_s}{\partial T} + \frac{\partial C_l}{\partial T})] \tag{5}$$

where H_0 is the net heating of the coupled ocean-atmosphere evaluated at a reference temperature T_0. $E = \sigma T^4$, where T is the SST, and σ is the Stefan-Boltzmann constant. G_a is the clear sky greenhouse effect, C_l is the greenhouse effect of clouds, and C_s is the cloud short-wave forcing, $\frac{\partial G_a}{\partial T} + \frac{\partial C_s}{\partial T} + \frac{\partial C_s}{\partial T}$ is the net radiative feedback of water vapor and clouds (3). ρ is the density, C_p is the specific heat, and h is the depth for the surface ocean. Equation (4) is obtained by linearizing the net heating of the coupled ocean-atmosphere about temperature T_0. In deriving (5), it is assumed that the atmosphere is always in thermal equilibrium with the underlying ocean. Equations (4) and (5) show that the more positive the radiative feedbacks in the atmosphere, the larger the T_e and the smaller the c, both of which lead to a larger α^*.

The equilibrium solutions for (1), (2) and (3) are shown in Figure 2. The coupled system contains two equilibrium states. For $\alpha^* < 1$ (weak coupling), the equilibrium is the radiative convective equilibrium, a warm state that has no ocean current, and it is a stable state. For $\alpha^* > 1$ (strong coupling), the radiative convective equilibrium becomes unstable, and a new state with an ocean current is switched on. The new state has a finite temperature difference between the east and the west, and is colder than the radiative-convective equilibrium.

Figure 2 shows that in the presence of large positive feedbacks from the atmosphere, the coupled system is able to drift automatically to a state with an ocean current. The ocean current transports heat from the surface ocean to the subsurface ocean, fighting against the positive feedbacks from the atmosphere that tend to warm the system.

Figure 2 further shows that once the system finds itself at the cold state with the presence of the ocean current, the larger the T_e, the larger the difference between T_e

and T_1. This is because the rate of increase of T_1 and T_2 with T_e decreases with the increase of T_e. It is easy to show that

$$\frac{dT_1}{dT_e} = \frac{1}{\sqrt{\alpha^*}}, \ \frac{dT_2}{dT_e} = \frac{1}{2\sqrt{\alpha^*}} \tag{6}$$

This leads to a very effective regulation on the temperature of the surface ocean. In fact, this regulatory effect may limit the temperature of the surface ocean to below 305 K even in the absence of cloud feedbacks. In the absence of feedbacks from clouds, c solely depends on water vapor feedback (equation(4). Using the water vapor feedback given by Sun and Oort (15), c is about $1.5 \times 10^{-8}s^{-1}$ Figure 3 plots T_1 and T_2 as a function of T_e with $T_c = 291K$ (18^oC) and $\alpha = 1.0 \times 10^{-8}K^{-1}s^{-1}$. The observed speed for the surface ocean current and SST gradients are used to estimate α. A direct estimate using the bulk drag formula leads to a similar value for α. We see that though T_1 and T_2 increase with T_e, the rate of increase is so small that they are practically independent of T_e. T_1 will be effectively limited to less than 305 K as long as T_e does not exceed 329 K. The maximum T_e estimated from equation (8 is about 324 K. This estimate was obtained by taking H_0 as the heating rate in a clear sky situation. The maximum value for T_1 and T_2 are also not sensitive to the strength of water vapor feedback. The value of c may become larger (smaller) for a weaker (stronger) water vapor feedback, but T_e will become smaller (larger)

To have a more realistic treatment of ocean dynamics, we have conducted experiments with an ocean general circulation model (GCM), the GFDL MOM model (16). The replacement of the three box model with an ocean GCM leads to very similar results. Parameterizing the heat exchange with the atmosphere the same way as in the box model, and coupling the east-west surface stress with the east-west SST difference, we again find that the ocean has two equilibrium states: a warm state that has no wind-driven currents and a cold state that has wind-driven currents. The warm state becomes unstable when the coupling between the wind stress and the east-west

SST difference is sufficiently strong. Figure 4 shows equilibrium values for the area averaged SST over the western and the eastern tropical ocean as a function of the coupling strength between the surface wind stress and the east-west SST differences. Clearly, the ocean GCM exhibits the same feature as the three box ocean model. Once the ocean finds itself in a state with wind-driven currents, the stronger the coupling, the colder the tropical ocean. With the presence of wind driven currents, the temperature difference between the east and the west remains almost unchanged, indicating the strong dynamic connection between these two regions.

The dynamic coupling between the western Pacific ocean and the eastern Pacific ocean, however, is completely ignored by Ramnathan and Collins (3). They over looked a simple, but important fact: it is the derivative of the oceanic transport with respect to SST, not the magnitude of the oceanic transport that determines the sensitivity of the tropical SST. The heat budget for an ocean-atmosphere column in an equilibrium can be written as follows,

$$H = S_c + C_s + G_a + C_l - E - F_o - F_{as} + F_{al} = 0 \qquad (7)$$

F_o and F_{as} are respectively the lateral heat transport by ocean currents and atmospheric circulations. F_{al} is the heating from a net moisture convergence. Other terms have the same meaning as in equation (4). Expanding H about a reference temperature T_0, we have

$$T \simeq T_0 + H_0[\frac{\partial F_o}{\partial T} + \frac{\partial E}{\partial T} - \frac{\partial G_a}{\partial T} + \frac{\partial F_{as}}{\partial T} - \frac{\partial C_l}{\partial T} + \frac{\partial F_{al}}{\partial T} - \frac{\partial C_s}{\partial T}]^{-1} \qquad (8)$$

where H_0 is the H evaluated at T_0, and δT is the temperature deviation from T_0. Ramanathan and Collins (3) dropped F_o from equation (7) before it is perturbed to obtain equation (8).

Equation (8) also shows that the effectiveness of the feedback from ocean currents depends crucially on the nature of other feedbacks. Over the warm-pool region, $\frac{\partial E}{\partial T}$

is largely cancelled by $\frac{\partial G_a}{\partial T}$, and $\frac{\partial C_l}{\partial T}$ is largely balanced by $\frac{\partial F_{as}}{\partial T}$ (3). There may be further cancellations between $\frac{\partial C_s}{\partial T}$ and $\frac{\partial F_{al}}{\partial T}$ since increased cloudiness over the warm pool region is very likely accompanied by enhanced moisture convergence (4). The large cancellations among atmospheric feedbacks makes the feedback from ocean currents much more important than what one might infer from the magnitude of the heat transport by the ocean currents alone. More importantly, all the feedback terms in equation (8)are not independent of the dynamic coupling that give rise to SST gradients and the accompnaying large-scale circulations. The increased cloudiness during the 1987 El Nino found by Ramanathan and Collins (3) is also a consequence of changes in east-west SST gradients and accompanying large circulations (4).

Figure 1: A schematic diagram for the coupled model.

Figure 2: Equilibrium solutions for the coupled model. (a): strength for the ocean current. Plotted is $f = \frac{q}{c}$. The two equilibrium solutions are $f = 0$ and $f = \sqrt{\alpha^*} - 1$. The latter solution exists only when $\alpha^* > 1$. (b): surface ocean temperatures. Plotted are $T_1^* = \frac{T_1 - T_e}{T_e - T_c}$ and $T_2^* = \frac{T_2 - T_e}{T_e - T_c}$. For $f = 0$, $T_1^* = T_2^* = 0$. For $f = \sqrt{\alpha^*} - 1$, $T_1^* = -1 + \frac{2}{\sqrt{\alpha^*}} - \frac{1}{\alpha^*}$ and $T_2^* = -1 + \frac{1}{\sqrt{\alpha^*}}$. The dashed line indicates that the solution exists, but is unstable.

Figure 3: T_1 and T_2 as a function of T_e for $c = 1.5 \times 10^{-8} s^{-1}$, $\alpha = 1.0 \times 10^{-8} K^{-1} s^{-1}$, and $T_c = 18^o C$. The maximum of T_e for $c = 1.5 \times 10^{-8} s^{-1}$ and the corresponding maximum surface ocean temperatures are marked in the figure.

Figure 4: The equilibrium SST in the western and the eastern tropical Pacific ocean as a function of the coupling strength. T_1 and T_2 are area averaged SST for the tropical western Pacific and tropical eastern Pacific respectively. λ_0 measures the sensitivity of wind stress to changes in SST gradients. The ocean model is the GFDL MOM model (Pacanowski et al 1991). The model domain is

$(0, 40^o) \times (2^o S, 50^o N) \times (0, 3000m)$ with a resolution of $2^o \times 2^o \times 15$ levels. With no wind imposed on the ocean, we first spin up the ocean GCM for 1000 years for the surface layer, and 5000 years for the bottom layer using an acceleration scheme given by Bryan (18). The heat exchange between the atmosphere and the ocean has the same form as in the box model (i.e., restoring SST boundary condition is used), with T_e varying with latitude following a cosine profile from 319 K at equator to 263 K at 50 N. The thermal relaxation time scale $\frac{1}{c}$ is chosen as 200 days. The salinity field is held constant. The spin up is to allow the thermohaline circulation to set up a basic temperature structure for the ocean. Except near the western boundary, SST of such a state is essentailly zonally symmetric. A perturbation to the upper ocean is introduced by imposing a weak wind stress for a year. The surface wind is then coupled to the east-west SST differences in a way similar to that for the box model: $\tau_x = \lambda_0 \eta(y)(T_1 - T_2)$, where τ_x is the east-west wind stress, $\eta(y)$ is a specified function of latitudes which gives easterly wind in the tropics and westerly wind in the extratropics. 13 experiments with different λ_0 are conducted. For small λ_0, no significant east-west SST gradients are developed, and the ocean returns to the basic state without wind. When λ_0 is sufficiently large, however, the ocean drifts quickly to a new state with significant east-west SST differences. No significant changes are found after the first couple decades of integration, consistent with the idea that the heat transfer is dominated by the wind-driven circulations. Plotted are values of T_1 and T_2 at the end of 50 years of integration. Note that α in equation (3) is proportional to λ_0, and thus varying λ_0 is equivalent to varying α^* when T_e and c are fixed.

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