Oceanic Regulation of the Atmospheric Walker Circulation

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ABSTRACT

A coupled theory is proposed to account for the magnitude of the Walker circulation in the tropical Pacific. It is suggested that the Pacific Walker circulation is at a saturation state, at which the zonal sea surface temperature difference is bounded by about a quarter of the latitudinal difference of the radiative–convective equilibrium sea surface temperature.

1. Introduction

A quarter of a century ago, Bjerknes (1969) proposed the existence of a zonal atmospheric circulation cell, the Walker circulation, that ascends in the western Pacific and subsides over the eastern Pacific. Today, the Walker circulation has been recognized to lie in the heart of several current climate issues. It plays a critical role in the El Niño–Southern Oscillation (ENSO) event (Philander 1990). Indeed, this cell was named the Walker circulation by Bjerknes because it was recognized to be an important part of Walker's Southern Oscillation, the atmospheric half of ENSO. Recently, the Walker circulation has also been recognized to play an important role in the regulation of the tropical sea surface temperature (SST) because of its efficient heat transport (Wallace 1992; Fu et al. 1993; Pierrehumbert 1995).

In a coupled ocean–atmosphere system, the couplet of the Walker circulation and the warm pool/cold tongue system can be initiated by the Hadley circulation under the forcing of the meridional differential heating, or by the zonal wind–upwelling positive feedback (McWilliams and Gent 1978; Neelin and Djikstra 1995). However, there has virtually been no theory that explains the magnitude of the Walker circulation. Given that the only external forcing to the coupled climate system is the solar radiation, which is zonally uniform, it is unclear why the observed zonal SST difference across the Pacific equator is about 5°C, which forces a Walker circulation with a surface westward wind stress of about 0.5 dyn cm^{-2}. Other fundamental issues related to the Walker circulation also remain to be explained. For example, the last glacial maximum (LGM) SST (CLIMAP Project Members 1976) seems to suggest an enhanced zonal SST difference, while the simulated future climate with increased CO_2 shows a significant reduction of the Walker circulation (Knutson and Manabe 1995). The reasons for the different changes of Walker circulation remain unclear. Furthermore, a recent intercomparison of a dozen fully coupled general circulation models (Mechoso et al. 1995) shows a surprisingly consistent zonal SST gradient along the equator among all the models, in spite of the dramatically different absolute SSTs (Fig. 1). Does this imply a saturation of the Pacific SST gradient, and in turn the Walker circulation, that is independent of the absolute equatorial SST?

Here we will give a brief overview of our recent work on the Walker circulation. We will show that the strength of the Walker circulation is indeed regulated below a saturation level. Furthermore, this regulation is caused by the ocean advection that tends to decouple the zonal SST gradient from the ocean

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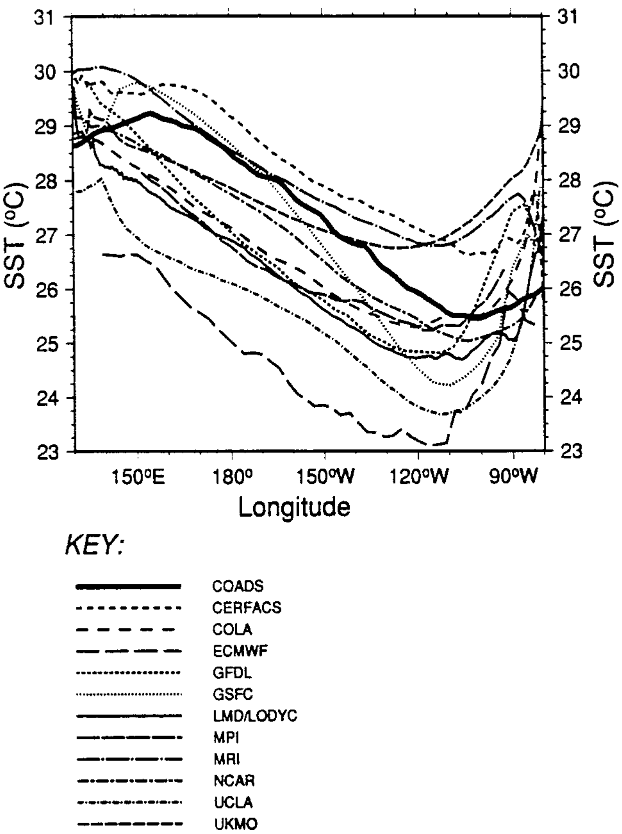

FIG. 1. Zonal SST profiles within 2° of the equatorial band for a dozen fully coupled GCMs without flux correction (adapted from Machoso et al. 1995). The most surprising feature is a consistent zonal SST gradient (except in the western and eastern margin where is affected by land). In contrast, the absolute equatorial SSTs differ dramatically among experiments.

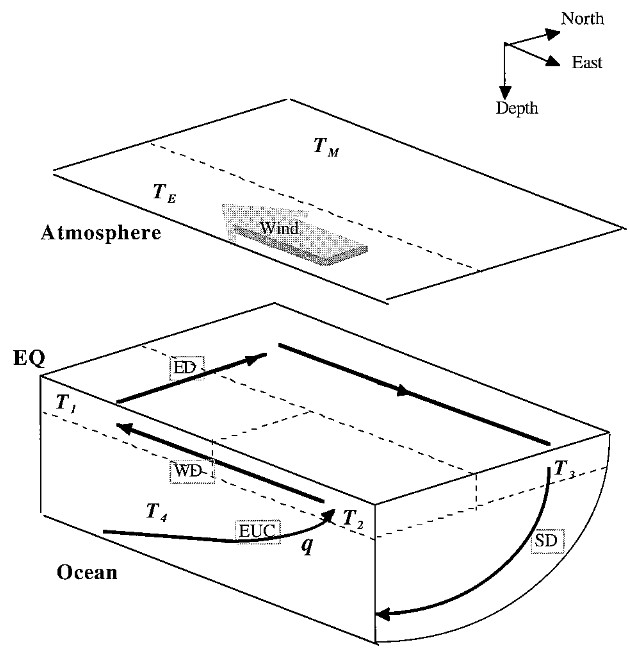

FIG. 2. Schematic figure for the four-box ocean model. The tropical–extratropical oceanic bridge consists of a net ocean transport, $q > 0$, upwelling into the eastern equator (box 2) through the Equatorial Undercurrent (EUC). The water then flows into the western equator (box 1) through the surface westward wind drift (WD) and diverges poleward into the midlatitude through the surface Ekman drift (ED). Finally, the water subducts (SD) toward the equator to maintain the equatorial thermocline, which provides the cold source water for upwelling.

currents. For simplicity, we will focus on the simplest possible model to highlight the physical mechanism, while a general theory and extensive numerical experiments will be given elsewhere (Liu and Huang 1996; Z. Liu 1997, manuscript submitted to *J. Climate*).

2. The coupled climate model

The ocean will be crudely simulated by four boxes (Fig. 2): boxes 1 and 2 for the surface western and eastern Pacific equator, respectively; box 3 for the surface midlatitude ocean; and box 4 for the combined tropical–extratropical subsurface thermocline water. The temperatures for each box are denoted by T_1, T_2, T_3, and T_4, respectively, while the surface heat fluxes into the three surface boxes are denoted by H_1, H_2, and H_3, respectively. One important feature of this model is a tropical–subtropical oceanic bridge associated with the Ekman flow as shown in Fig. 2 (Pedlosky 1987; Liu et al. 1994; McCreary and Lu 1994). The two equatorial boxes, assumed to have equal volumes, are controlled by the heat equations

$$\frac{d}{dt}T_1 = H_1 + q(T_2 - T_1),$$
$$\frac{d}{dt}T_2 = H_2 + q(T_4 - T_2), \tag{1}$$

where q represents the transport of the ocean current.

The surface heat flux is determined by the local air–sea thermal coupling, which can be crudely approximated as a restoring toward a radiative–convective equilibrium temperature with a relaxation time of τ. With atmospheric feedbacks and for large-scale flows, τ may reach several years for an oceanic mixed layer of 50 m (Bretherton 1982; Sun and Liu 1996). Here the equatorial SSTs (T_1 and T_2) will be restored toward an equatorial equilibrium temperature T_E, while the midlatitude SST (T_3) will be restored toward a midlatitude equilibrium temperature T_M ($< T_E$). That is,

$$H_1 = (T_E - T_1)/\tau,$$
$$H_2 = (T_E - T_2)/\tau, \tag{2}$$

and $H_3 = (T_M - T_3)/\tau$. The two equilibrium temperatures represent the only external forcing to the coupled ocean–atmosphere system, which warms the Tropics more than the midlatitude.

The ocean upwelling transport q is determined by the dynamic ocean–atmosphere coupling because it is driven by surface wind that in turn is forced by the SST gradient. Assuming the upwelling to be caused by the Walker circulation alone for the time being, q is proportional to the zonal wind and, in turn, the zonal SST gradient as

$$q = a(T_1 - T_2) \geq 0, \tag{3}$$

with a representing the dynamic coupling strength.

For simplicity here, boxes 3 and 4 will be assumed infinitely large and small, respectively. The heat equations for boxes 3 and 4 are then reduced to $T_3 = T_4 = T_M$. Then the ocean model (1), when combined with the thermal coupling (2) and dynamic coupling (3), forms a coupled ocean–atmosphere system in which the only external forcing is the latitudinal differential heating $T_E - T_M$. The steady-state solution will be determined by a single nondimensional coupling parameter, $A = a(T_E - T_M)$, which can be shown proportional to the ratio of times between the local relaxation and advection.

3. Regulation of Walker circulation

For weak coupling $A \leq 1$, there is a single stable state, $q = T_1 - T_2 = 0$, representing the local equilibrium state without Walker circulation. For strong coupling $A > 1$, this local equilibrium is destabilized by the Bjerknes wind/upwelling positive feedback (or the climatological version of Bjerknes positive feedback by Neelin and Djikstra). A new stable state emerges with $q\tau = \sqrt{A} - 1 > 0$, and $T_1 - T_2 = (T_E - T_M)(\sqrt{A} - 1)/A > 0$. This state has a finite transport, zonal SST gradient, and Walker circulation (Fig. 3a).

The most surprising feature in Fig. 3a is that, at modest coupling $A = 4$, the Walker circulation (solid) achieves its maximum, or saturation level:

$$(T_1 - T_2)_{\mathrm{Sat}} = (T_E - T_M)/4. \tag{4}$$

It is important to realize that this maximum amounts to only a quarter of the trivial upper bound of the coupled system $T_E - T_M$. Thus, an increase in the coupling strength enhances the Walker circulation only for weak coupling of $A < 4$. With stronger coupling with $A > 4$, the Walker circulation will no longer intensify. Hence, the Walker circulation is regulated by an upper bound that corresponds to a zonal SST difference about a quarter of the latitudinal difference of the equilibrium SST. For a local relaxation time, $\tau = 200$ days, typical values in the Pacific give $A \approx 4$, which is strong enough to produce a saturated Walker circulation. Furthermore, a reasonable value for $T_E - T_M$ is about 20°C. This will put the upper bound for zonal SST difference at about 4°–5°C, comparing well with the observed Pacific.

The saturation of Walker circulation seems surprising at first. One might expect an increased coupling to produce a stronger upwelling because of an increased efficiency to generate ocean currents. This is indeed the case in the model because the transport increases monotonically with A (dashed curve in

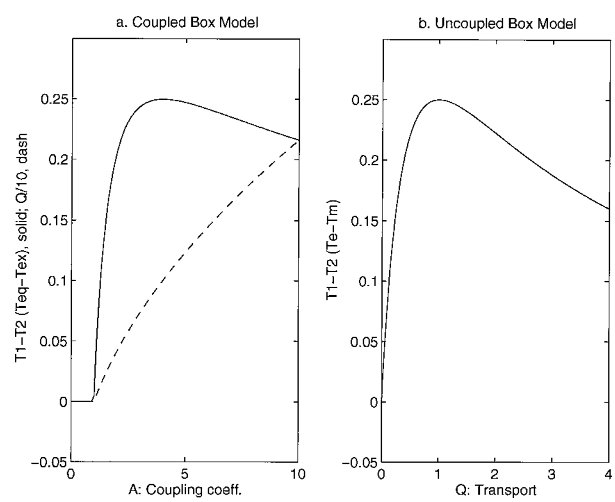

FIG. 3. Box model solutions. (a) Stable steady states for the coupled model. The zonal SST difference $T_1 - T_2$ (or the Walker circulation) (solid) and the transport $Q \equiv q\tau$ (dash) as function of A. (b) Uncoupled ocean-alone model with a prescribed transport Q; the SST difference $(T_1 - T_2) = (T_E - T_M)Q/(Q^2 + 2Q + 1)$ is plotted as a function of Q.

Fig. 3a). One might further expect a stronger upwelling to increase the zonal SST gradient. Now there is a caveat. If the current is so strong that the local relaxation becomes negligible, the zonal SST field will become almost uniform downstream. In other words, with too strong currents, the downstream SST gradient will be decoupled from the current strength purely due to ocean advection. A maximum Walker circulation is achieved when the timescale of advection is comparable to that of the local equilibration.

The decoupling of the SST gradient from strong ocean currents seems to suggest that the regulation arises purely from the ocean dynamics in the coupled system or independently of dynamic coupling. To confirm this, the uncoupled ocean-alone model that consists of (1) and (2) (and $T_3 = T_4 = T_M$) is used. This model is essentially the same as the coupled model, except for a given wind-driven current transport q [as opposed to a current that is caused by the dynamic coupling (3)]. The steady-state SST difference is plotted against ocean transport in Fig. 3b. Clearly, the SST difference increases with q for $q < 1$ but decreases with q for $q > 1$. Thus it is concluded that the regulation of Walker circulation in the coupled model is caused by the ocean dynamics that decouple the zonal SST gradient from the surface wind-drift current in the presence of strong currents.

4. GCM experiments

Extensive experiments with the Geophysical Fluid Dynamics Laboratory ocean general circulation model (Pacanowski et al. 1991) have been performed to confirm the results of the box model. Here, we only present two sets of experiments: one coupled and the other uncoupled. The coupled model is a hybrid coupled GCM, in which the ocean model is coupled with a simple atmosphere. The atmospheric zonal wind stress is assumed to have a fixed spatial pattern that varies with latitude only (the meridional wind stress is set to zero). The magnitude of the zonal wind stress, however, depends on the equatorial zonal SST difference and therefore gives the interactive coupling.

In each coupled experiment, an initial SST anomaly is added, and the model is allowed to evolve in the coupled mode for 100 yr when the upper ocean has reached quasi-equilibrium. Twelve coupled experiments with increasing coupling strength are presented in Fig. 4a for the zonal SST difference (solid) at final equilibrium. The solution resembles remarkably well the box model solution in Fig. 3a. For weak coupling, the SST relaxes to the initial state with little zonal SST difference. With strong coupling, the solution jumps to a state with finite zonal SST difference. Most importantly, for a further increase of the coupling, the zonal SST difference remains virtually unchanged at a saturation level of about 4°C, which is about one-quarter of the tropical–extratropical difference of restoring temperature. In contrast, the transport at the bottom of the eastern Pacific equator (dashed curve in Fig. 4a) increases monotonically with the coupling, agreeing with the box model solution in Fig. 3a. Thus the GCM experiments confirm the saturation of Walker circulation.

Furthermore, to confirm the effect of ocean dynamics on the regulation of Walker circulation, nine uncoupled ocean-alone experiments are carried out with increased wind forcing. The result (Fig. 4b) again agrees with the box model solution in Fig. 3b. The most important feature is an increase of SST gradient with the wind for weak winds but a saturation of SST gradient for strong winds.

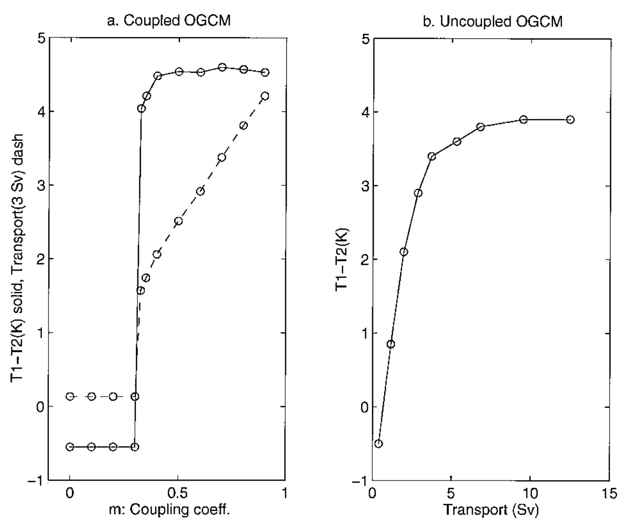

FIG. 4. MOM results. (a) Twelve coupled experiments, in which the intensity of the zonal wind stress depends on the SST as $\tau^x = m(T_1 - T_2)M(\theta)$. The $T_1 - T_2$ (solid) and the upwelling transport into the eastern equator (dashed) are plotted against the coupling parameter m. (b) Nine uncoupled ocean-alone experiments in which the wind is given by $\tau^x = dM(\theta)$, with d values of 0, 0.25, 0.5, 0.75, 1, 2, 3, 4. The $T_1 - T_2$ is plotted against the transport for each run. The solutions in (a) and (b) resemble remarkably well the box model results in Figs. 2a and 2b, respectively. Here, T_1 and T_2 are, respectively, the SSTs averaged in the western and eastern thirds of the equatorial ocean within the latitude of 10°. The wind pattern M(θ) is fixed with easterly wind in the Tropics and westerlies in the midlatitude.

5. Discussion

It has been shown that the strength of the Walker circulation is regulated by an upper bound that depends on the latitudinal differential heating as in (4). The regulation is caused by strong ocean advection that decouples zonal SST gradient from ocean currents. One can further show that this saturation level will not be changed after the inclusion of the coupling with the Hadley circulation (Liu and Huang 1996). Indeed, as shown in the ocean-alone model, this saturation level is independent of the dynamic coupling between the ocean and atmosphere. As long as the wind is strong enough, no matter if it is caused by the Walker or Hadley circulation, the saturation will be achieved.

This nontrivial saturation level does not exist in previous works because of the neglect of either the interactive surface heat flux or the zonal advection along the equator. Indeed, it can be shown that for a fixed heat flux (as in McWilliams and Gent 1978) the zonal SST difference can become infinite, while in the absence of the zonal advection (as in Neelin and Djikstra 1995) the upper bound for the zonal SST difference increases to the trivial upper bound $T_E - T_M$.

The saturation seems to explain the present Pacific very well as discussed above. It may also explain the change of tropical climate for the LGM and the CO_2 warming experiments. The cooling of the LGM climate has a polar amplification, with the midlatitude cooled more than the equator. This increases the latitudinal differential equilibrium SST and in turn the saturation level of the Walker circulation according to (4). The opposite occurs for the CO_2 warming experiment. Furthermore, the consistent zonal SST gradient in the fully coupled GCM experiments in Fig. 1 can be viewed as an evidence of the saturation of Pacific zonal SST gradient and Walker circulation in fully coupled models.

One direct application of the theory is the regulation of the tropical SST (Sun and Liu 1996). It is well known that the ocean dynamic upwelling is important for the cooling of the eastern Pacific cold tongue. If the Pacific is indeed in the saturation state, the western warm pool SST should be locked with the eastern cold tongue at long timescales. Thus, although our theory can not explain the absolute regulation of the tropical SST, it does suggest that the warm pool SST can be directly regulated by the cold tongue through ocean dynamics.

As a final note, the solutions that we discussed above are equilibrium solutions, which are valid after the decadal timescale adjustment of the subtropical thermocline ventilation process. Thus, our theory does not apply to short timescale climate variability such as the ENSO. Indeed, at short timescales, recent studies show that the zonal SST gradient may exhibit different response from our equilibrium solution (Clement et al. 1996; Seager and Murtugudde 1997; Z. Liu 1997, manuscript submitted to *J. Climate*).

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