Coordinated heat removal from the equatorial Pacific during the 1986-87 El Niño

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Abstract. Utilizing radiation data from Earth Radiation Budget Experiment (ERBE), circulation statistics from NCEP reanalysis, and assimilated ocean data for the tropical Pacific basin, we show that the surface ocean warming during the 1986-87 El Niño is not only accompanied by significant increases in the cloud reflection of the solar radiation, but also by marked increases in the poleward energy transport in both the atmosphere and ocean. Measured over the equatorial region, the feedback from the ocean dynamics is twice as large as from the atmospheric dynamics which in turn is twice as large as the feedback from the cloud albedo. The three feedbacks constitute a strong regulatory effect upon the equatorial SST. The results reveal a prominent role of El Niño in the heat removal from the equatorial Pacific.

Introduction

The surface Pacific Ocean warming during El Niño results in feedbacks from clouds [Ramanathan and Collins 1991], atmospheric circulation [Bjerknes, 1966; Oort and Yienger, 1996], and ocean dynamics [Wyrtki, 1985]. The last two feedbacks, however, have not been quantified from observations. Using the best data that are currently available, we attempt to quantify the radiative and dynamical feedbacks in an integrated framework and to obtain a better view of the role of El Niño in the heat removal from the equatorial Pacific.

Ocean warming during El Niño is mostly confined in the equatorial Pacific region where the atmosphere is strongly coupled with the ocean [Philander, 1990]. The energy budget for a column ocean-atmosphere in thermal equilibrium may be written as

$$H = S - F + D_a + D_o = 0 \qquad (1)$$

where H is the net energy flux into the coupled ocean-atmosphere column, S and F are respectively the net downward solar radiation and the net outgoing long-wave radiation at the top of the atmosphere, D_a is the convergence of moist static energy in the atmosphere, and D_o is convergence of heat in the ocean [Trenberth and Solomon, 1994] (D for "dynamics" including the effects of mean circulation as well as transients). Further expanding S and F, we obtain

$$H = S_c + C_s - \sigma T^4 + G + D_a + D_o = 0 \qquad (2)$$

where S_c is the clear sky solar radiation, C_s is the shortwave cloud forcing, σT^4 is the surface emission (T is the SST and σ is Stefan-Boltzman constant), and G is the sum of greenhouse effect of water vapor (G_a) and clouds C_l) [Ramanathan and Collins, 1991] ($G = \sigma T^4 - F$). Linearizing Eq. (2) about a reference temperature T_0 results in the equation for the equilibrium SST,

$$T \simeq T_0 + H(T_0)/\beta \qquad (3)$$

where T is the equilibrium SST, and β is given by

$$\beta = 4\sigma {T_0}^3 - \frac{\delta G}{\delta T} - \frac{\delta C_s}{\delta T} - \frac{\delta D_a}{\delta T} - \frac{\delta D_o}{\delta T}. \qquad (4)$$

Thus, the equilibrium SST in the tropics is determined by both radiative and dynamical feedbacks. Recognizing that oceanic warming during El Niño is a basin-wide coupled phenomena which involves zonal redistribution of water in the equatorial upper ocean, we estimate all the feedback terms in the above equation in a single framework using the best information currently available. The radiative terms have been assessed by Ramanathan and Collins [1991], but their analysis did not fully take into account the basin-wide coupled nature of El Niño. Consequently, they did not thoroughly separate geographical from temporal variations in their calculations of the relationship between changes in radiative fluxes and changes in SST.

Heat removal by dynamics and clouds

We start with the assessment of the relative role of D_a and C_s by computing the differences in SST, D_a, and C_s between April 1987 (an El Niño year) and April 1985 (a normal year) (Fig. 1). SSTs are from the standard AMIP SST data set at T42 resolution [Gates 1992], C_s is from ERBE [Barkstrom et al., 1990], and D_a is calculated from NCEP reanalysis [Trenberth and Solomon, 1994; Trenberth and Guillemot 1998]. Changes in D_a have a similar spatial pattern to changes in C_s, and both tend to damp the SST anomaly. In regions with maximum warming the magnitude of changes in D_a is twice as large as that of changes in C_s (the color scale for D_a (Fig. 1) is two times of that for C_s). There is a mean poleward transport of moist static energy from the tropical Pacific [Trenberth and Solomon, 1994]. Accordingly, the spatial pattern of changes in D_a indicates an enhanced poleward transport of moist static energy during El Niño.

The time series of SST, D_a, and C_s averaged over the equatorial belt (5^oS to 5^oN and zonally across the entire Pacific Ocean (120^oE to 70^oW with the land areas excluded) convey the same information as Fig. 1 (see Fig. 2): there are marked increases in both C_s and D_a from the cool phase to the warm El Niño phase. Relative to the climatological mean, there is enhanced cloudiness over that region from mid-1986 to mid-1987, accompanied by enhanced poleward divergence of energy in the atmosphere that begins as SST anomalies rise late in 1985 . The magnitude of the variations in D_a in the equatorial belt is larger and more sustained than changes in C_s. This result is insensitive to the width of meridional belt over which D_a and C_s are averaged as long as it is limited to the regions where significant ocean warming takes place.

For the equatorial region, the change in the net radiative flux at the top of the atmosphere ($S-F$) during El Niño is observed to be small (less than about 5 Wm^{-2}), and consequently the energy flux divergence in the atmosphere is mainly provided by changes in the surface heat flux into the ocean F_s (Fig. 3a). F_s was obtained as a residual by considering the energy balance of the atmosphere,

$$F_s = S - F + D_a - \frac{\partial E_a}{\partial t} \tag{5}$$

where $\frac{\partial E_a}{\partial t}$ is the change in storage in the atmosphere of the total energy (the sum of the moist static energy and the kinetic energy) [Trenberth and Solomon, 1994]. S and F were computed from ERBE data. $\frac{\partial E_a}{\partial t}$ was obtained from NCEP reanalysis. The heat flux into the ocean is significantly reduced as the El Niño warming develops, consistent with early findings [Barnett et al., 1991].

The El Niño warming results in even larger changes in the divergence of heat within the ocean (Fig. 3b). D_o was calculated as a residual from the following equation

$$D_o = \frac{\partial H_c}{\partial t} - F_s \tag{6}$$

where H_c is the heat content of the upper ocean. H_c was obtained from the tropical Pacific Ocean analyses prepared by NCEP using an ocean data assimilation system [Ji et al., 1995]. The ocean temperatures from this data set have been shown to be accurate in representing annual and interannual variability [Ji et al., 1995; Smith and Chelliah, 1995]. The depth of the upper ocean is chosen as 380 m. Below this depth, the variations in the ocean temperature field are small (less than 0.1 K). Because the variations of temperature below the upper ocean are small, D_o mainly represents the horizontal convergence of heat in the ocean, and thus there is an enhancement of the normal poleward transport of heat in the ocean [Trenberth and Solomon, 1994] during the 1986-87 El Niño. The oceanic transport plays a bigger role than clouds and the atmospheric transport in opposing the rise in the tropical SST.

Using time series of SST, C_s, D_a (Fig. 2) and D_o (Fig. 3), we can further quantify the feedback from cloud albedo, atmospheric transport, and oceanic transport through linear regression (Table 1). Other feedbacks in Eq. (4) were obtained the same way. The magnitude of cloud feedbacks is smaller than estimated by Ramanathan and Collins [1991] because a zonal average across the ocean is involved in order to be consistent with the estimate of the feedback from the oceanic transport. The other difference is that in our calculation, we used data over the entire ERBE period, thus minimizing the sampling errors that can be significant if the differences between just single months are used in the calculations. The error bars were obtained using the method of Press et al. [1992] for the case in which the measurement errors of individual data points are not known. A more thorough error analysis will be given in an extended paper. With the use of the feedbacks in Table 1 and taking $T_0 = 300$ K and $H(T_0) = 76\ W/m^2$ [Ramanathan and Collins, 1991], we estimate from Eqs. (3) and (4) that the maximum value for the zonally averaged equatorial Pacific SST is about 302 K, which is in good agreement with the observed peak values.

Discussion

The increase in the cloud reflection of solar radiation and the enhancement in the poleward energy transport in both the atmosphere and ocean are a direct consequence of the displacement of warm water from the west to the east. More deep convection takes place in the central Pacific because of the presence of warm water from the warm-pool region which increases the SST gradients between the subtropics and the deep tropics and thereby enhances moisture convergence to that region. The temperature of the surface water exceeds the threshold value for the onset of deep convection. The corresponding increase in the latent heat released in the equatorial region and a more zonal distribution of this heating drive a more vigorous meridional overturning in the atmosphere (i.e., the Hadley circulation) (Oort and

Yienger, 1996), which transport more energy away from the equatorial region in the atmosphere. The presence of warm water in the central and eastern Pacific reduces the equatorward heat transport by the geostrophic flow and increases the poleward heat transport by the Ekman divergence [Brady, 1994]. The disappearance of the tropical instability waves in the eastern Pacific during the El Niño warming may have also contributed to the anomalous poleward heat transport in the ocean since these waves normally transport a significant amount of heat equatorward [Legeckis, 1977; Hansen and Paul, 1984]. Another consequence of this loss of warm water is that the relatively cold thermocline below the warm-pool shoals which enhances the cooling effects of the equatorial upwelling upon the warm surface ocean [Philander, 1990].

It is apparent from the above analysis that the occurrence of El Niño is an effective way to remove heat from the equatorial Pacific. Clouds, atmospheric circulation, and ocean dynamics appear to work in a fairly coordinated manner during this process in counteracting the effects of solar heating and the greenhouse effect. This raises the question of whether the very existence of El Niño arises from the need to move heat out of the equatorial Pacific. Extending recent studies on the regulation of the tropical SST [Sun and Liu, 1996; Clement et al., 1996; Seager and Murtugudde 1996], Sun [1997] found in a simple coupled model of the equatorial Pacific that El Niño only exists when the intensity of the radiative heating in the equatorial region (or the SST of the warm-pool) exceeds a critical value. With the occurrence of El Niño, the heat removal in the equatorial Pacific region is largely divided into two stages. The first stage is to store the heat from the solar radiation and the greenhouse effect into the upper ocean. During this stage, relatively cold surface water upwelled from the central and eastern equatorial Pacific absorbs heat from the sun and the greenhouse effect during its passage to the western side of the basin. The second stage is the redistribution of warm water and the depletion of the upper ocean heat content; a stage with highly coordinated heat removal from the equatorial Pacific (Figs. 1, 2, 3). Thus the need to remove heat from the equatorial Pacific appears to be fundamental to the existence of El Niño.

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Figure 2. Variations of SST, C_s, and D_a over the period of ERBE. Plotted are their monthly anomalies averaged over the entire equatorial Pacific (5^{o}S-5^{o}N, 120^{o}E-70^{o}W). The corresponding time mean values for C_s and D_a are respectively –48 Wm^{-2} and –5.0 Wm^{-2}. During a normal year, there is a large cancellation between the divergence of moist static energy over the warm-pool and the convergence of moist static energy over the cold-tongue.

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Figure 3. (a): Variations of $S - F$ and F_s over the period of ERBE. Plotted are monthly anomalies averaged over the equatorial Pacific ocean (120^{o}E-70^{o}W). The corresponding time mean values of $S - F$ and F_s over the ERBE period are respectively 73 Wm^{-2} and 68 Wm^{-2}. (b): Variations of tendency of equatorial upper ocean heat content $\frac{\partial H_c}{\partial t}$ and convergence of heat into the equatorial upper ocean. H_c was calculated from monthly mean ocean temperatures for the upper 380 m. The variations of temperature in the upper 200 m account for about 90% of the total variability in H_c. After removing the annual cycle based on the climatology for the ERBE period, we used a Hanning window (a cosine bell window) with a width of seven months to eliminate high frequency noise in H_c. This smoothing reduces the total variance of H_c by about 8%. A central difference scheme was then employed to obtain the time derivative of H_c. The equatorial ocean is defined here as the 5^{o}S-5^{o}N equatorial belt spanning from 120^{o}E to 70^{o}W. The time mean value of D_o is –68 Wm^{-2}.

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Table 1. Radiative and Dynamical Feedbacks

Name of the Process	Feedback($Wm^{-2}K^{-1}$)	Correlation
$\frac{\delta C_s}{\delta T}$	-6.2 ± 1.0	-0.67
$\frac{\delta D_a}{\delta T}$	-11.1 ± 1.3	-0.77
$\frac{\delta D_o}{\delta T}$	-25.7 ± 3.8	-0.69
$\frac{\delta G}{\delta T}$	14.6 ± 0.9	0.92
$\frac{\delta C_l}{\delta T}$	8.1 ± 0.8	0.83
$\frac{\delta G_a}{\delta T}$	6.5 ± 0.2	0.97

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