Role of Climate Feedback in El Niño–Like SST Response to Global Warming

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ABSTRACT

Under global warming from the doubling of CO₂, the equatorial Pacific experiences an El Niño–like warming, as simulated by most global climate models. A new climate feedback and response analysis method (CFRAM) is applied to 10 years of hourly output of the slab ocean model (SOM) version of the NCAR Community Climate System Model, version 3.0, (CCSM3-SOM) to determine the processes responsible for this warming. Unlike the traditional surface heat budget analysis, the CFRAM can explicitly quantify the contributions of each radiative climate feedback and of each physical and dynamical process of a GCM to temperature changes. The mean bias in the sum of partial SST changes due to each feedback derived with CFRAM in the tropical Pacific is negligible (0.5%) compared to the mean SST change from the CCSM3-SOM simulations, with a spatial pattern correlation of 0.97 between the two. The analysis shows that the factors contributing to the El Niño–like SST warming in the central Pacific are different from those in the eastern Pacific. In the central Pacific, the largest contributor to El Niño–like SST warming is dynamical advection, followed by PBL diffusion, water vapor feedback, and surface evaporation. In contrast, in the eastern Pacific the dominant contributor to El Niño–like SST warming is cloud feedback, with water vapor feedback further amplifying the warming.

1. Introduction

The equatorial Pacific sea surface temperature (SST) changes associated with natural climate variability [e.g., El Niño–Southern Oscillation (ENSO)] affect not only tropical cyclone activity, freshwater supplies, agriculture, and ecosystems, but also the global patterns of flood and drought and climate extremes worldwide through atmospheric teleconnections (see review by McPhaden et al. 2006; Alexander et al. 2002). This indicates that the changes in equatorial Pacific SST are highly influential for global climate. Therefore, an accurate projection of the equatorial Pacific SST in response to increased greenhouse gases is fundamentally important to improving the projection of future global and regional climate change. This is not only because the changes in the background state of the equatorial Pacific may influence the regional climate in many parts of the world through atmospheric teleconnections, as illustrated by observed climate variability, but also because they could influence the amplitude, frequency, spatial pattern, and impacts of ENSO (Timmermann et al. 1999; Fedorov and Philander 2000; Koutavas et al. 2002; Vecchi and Wittenberg 2010; Collins et al. 2010; McGregor et al. 2013).

How will the equatorial Pacific respond to increased greenhouse gases? Will the pattern of SST changes more closely resemble an El Niño or a La Niña? This has been debated for over a decade (Vecchi et al. 2008). Although the majority of global climate models (GCMs) exhibit an El Niño–like warming in the equatorial Pacific, namely stronger warming in the central and eastern equatorial Pacific than in the western equatorial Pacific (Knutson and Manabe 1995; Meehl and Washington 1996; Cubasch et al. 2001; Yu and Boer 2002; Meehl et al. 2007; Vecchi and Soden 2007; An et al. 2012; Yeh et al. 2012; Lee and Wang 2014), there are also models that project a La Niña–like warming, with the western equatorial Pacific SST warming more than the central and eastern Pacific (Liu et al. 2005; An et al. 2012), or no significant trend toward either El Niño–or La Niña–like conditions but an enhanced equatorial warming relative to the subtropics (Collins 2005; Liu et al. 2005; DiNezio et al. 2009). Note that some recent studies (DiNezio et al. 2010; Collins et al. 2010; Xie et al. 2010) suggested...
that “El Niño (La Niña)–like climate change” is not an appropriate expression because of a number of differences between ENSO events and the tropical Pacific response to global warming, such as thermocline change, seasonality, and atmospheric teleconnections, among others. In this study, we use the term “El Niño–like warming” only to denote the pattern of mean SST change in the equatorial Pacific, which resembles a present-day El Niño event. Currently it is still a challenge to accurately determine the twentieth-century trend in the zonal SST gradient across the equatorial Pacific from observations/reconstructions because of uncertainties related to changing measurement techniques and analysis procedures (Deser et al. 2010). Recently Tokinaga et al. (2012a) reconstructed the SST trend pattern for the period 1950–2009 using only bucket measurements from the International Comprehensive Ocean–Atmosphere Data Set (ICOADS) and nighttime marine surface air temperature. They found that the zonal SST gradient between western and eastern Pacific is reduced, physically consistent with subsurface temperature observations.

Several mechanisms have been proposed to explain the formation of El Niño–like warming in response to increased greenhouse gases. First, the weakening of the Walker circulation and associated Bjerknes (1969) feedback was suggested (Meehl and Washington 1996; Yu and Boer 2002). Observational (Zhang and Song 2006; Vecchi et al. 2006; Tokinaga et al. 2012b), theoretical (Held and Soden 2006; Tokinaga et al. 2012a; Knutson and Manabe 1995), and modeling work (Vecchi and Soden 2007; Meehl et al. 2007) all indicate a weakening of the Walker circulation in response to global warming. The associated weakening of the equatorial trade winds may lead to a reduction in the upwelling of cold water in the eastern Pacific, resulting in a decrease of zonal SST gradient across the equatorial Pacific, which in turn may further weaken the Walker circulation and trade winds (Vecchi and Soden 2007; Meehl et al. 2007). The second mechanism is evaporation cooling (Knutson and Manabe 1995). Because of the nonlinear dependence of saturation mixing ratio on temperature through the Clausius–Clapeyron relation, the sea surface with higher temperature would produce more evaporative cooling in response to the same temperature perturbation. Thus, there will be stronger evaporative cooling in the western Pacific warm pool than the eastern Pacific cold tongue, which can damp the zonal SST gradient and result in El Niño–like warming (Knutson and Manabe 1995). This evaporative cooling argument was first proposed by Newell (1979) to explain the regulation of the warm pool SST. However, it is not supported by observations, which show low surface evaporation in the western Pacific warm pool region (Zhang and McPhaden 1995; Zhang et al. 1995; Zhang and Grossman 1996). Meehl and Washington (1996) showed stronger evaporative cooling in the eastern Pacific than in the western Pacific in a GCM due to larger SST increase in the east. In other words, evaporative cooling may act to enhance the zonal SST gradient instead of damping it. The third mechanism is cloud feedback. Ramanathan and Collins (1991) suggested that the increased highly reflective cirrus cloud associated with enhanced deep convection over the western Pacific warm pool as SST increases can block out more incoming solar radiation. This negative shortwave cloud feedback over the warm pool can lead to a reduction in east–west SST gradient across the equatorial Pacific. In addition, the decrease in low-level stratus clouds in the eastern Pacific as SSTs increase allows more incoming solar radiation to reach the surface. This positive cloud feedback in the eastern Pacific can also contribute to the reduction in zonal SST gradient across the equatorial Pacific (Meehl and Washington 1996). However, Yu and Boer (2002) showed that both El Niño–like warming and negative shortwave cloud feedback in central and eastern Pacific occur in the Canadian Centre for Climate Modelling and Analysis (CCCma) coupled GCM, which led them to conclude that cloud feedback is not a significant factor in the El Niño–like warming response.

The role of ocean dynamical processes has also been investigated. The SST is primarily regulated by ocean dynamical thermostat (cold water upwelling) in the eastern Pacific cold tongue (Clement et al. 1996; Cane et al. 1997). Since the increased downward radiative flux associated with a doubling of CO2 in the cold tongue region may enhance the vertical ocean temperature gradient and hence cold water upwelling, the weaker warming in the eastern Pacific than in the western Pacific will strengthen the east–west SST gradient across the equatorial Pacific, which drives stronger Walker circulation and surface easterly winds, and then results in stronger upwelling cooling in the eastern Pacific. This in turn further reinforces the zonal SST gradient and Walker circulation, indicating that the ocean dynamical process acts to enhance the zonal SST gradient. This ocean-dynamics-triggered Bjerknes feedback has been used to explain how a La Niña–like response can occur (Cane et al. 1997; Vecchi et al. 2008). The comparison of the tropical Pacific response to increased CO2 in 13 GCMs that participated in phase 3 of the Coupled Model Intercomparison Project (CMIP3) of the World Climate Research Programme shows that when the model uses a nondynamic slab ocean component, the equatorial Pacific exhibits an El Niño–like warming response, whereas when a full dynamical ocean component is used in the same model, the “El Niño-ness” of
the SST response is diminished (Vecchi et al. 2008). This demonstrates that the ocean dynamics indeed acts to strengthen the east–west SST gradient and thus damps the El Niño–like response, suggesting that atmospheric processes are the drivers of the formation of El Niño–like warming. Therefore, although equatorial Pacific SST variability is largely influenced by ocean dynamics, the slab ocean model version of climate model is still a good tool to help us understand the responsible mechanisms of the El Niño–like SST warming.

Because the Bjerknes feedback is excluded in a non-dynamic slab ocean model configuration, the appearance of an El Niño–like warming response suggests that the Bjerknes feedback from a weakened Walker circulation is not a dominant mechanism for such a warming response. The role of cloud feedback in most mechanism analyses is determined by cloud radiative forcing. However, the changes in noncloud climate components may affect (or contaminate) the change in cloud radiative forcing, which may even change the sign of global mean cloud feedback (Shell et al. 2008; Soden et al. 2008). Therefore, the contribution of cloud feedback to El Niño–like warming is still not very clear. Previous studies suggest that the ocean surface evaporation can modulate the east–west SST gradient in the equatorial Pacific. The changes in surface evaporation and circulation may affect the moisture content in the atmosphere. Can the associated water vapor feedback influence the SST response in the equatorial Pacific? In addition, when the Walker circulation is weakened, the intensity and location of convection will change correspondingly. Will the changes in temperature lapse rate affect the zonal gradient of surface warming in the equatorial Pacific? All these questions indicate that our understanding on the mechanisms that govern the equatorial SST response to increased greenhouse gases is still incomplete.

The SST changes are usually analyzed using the heat budget equation of ocean mixed layer. Although the contributions of surface turbulent and radiative heat fluxes and ocean heat transport can be explicitly determined, this method cannot isolate the contributions of each radiative climate feedback (e.g., water vapor feedback and lapse rate feedback) or the contributions of each physical and dynamical process of GCMs. The traditional climate feedback analysis methods (e.g., partial radiative perturbation method; Wetherald and Manabe 1988) focus on the feedback parameter, the ratio of the radiative perturbation at the top of the atmosphere due to a specific feedback to the total change of the surface temperature, and do not provide a direct estimate of surface temperature change caused by any individual feedback. Therefore, it is difficult to directly use these methods to quantify the contribution of climate feedback to SST changes. Recently, Lu and Cai (2009) formulated a coupled atmosphere–surface climate feedback and response analysis method (CFRAM). This method can explicitly quantify the contributions of both radiative and nonradiative climate feedbacks to temperature changes. The decomposition of nonradiative feedback is based on the GCM’s physical and dynamical processes so that the roles of GCM physical and dynamical processes in climate change can be easily understood. Song et al. (2014) applied the CFRAM to 10 years of hourly output of the slab ocean model (SOM) version of the National Center of Atmospheric Research (NCAR) Community Climate System Model, version 3.0, (CCSM3-SOM) to quantify the contributions of climate feedbacks to global pattern of surface warming with particular interests in the warming contrast between land and ocean and between high and low latitudes in response to a doubling of CO₂. They found that the CFRAM analysis with hourly model output is highly accurate for quantifying the role of climate feedbacks in the spatial distribution of global and regional warming projected by GCMs. In this study, we use the same method and data to understand the equatorial Pacific SST response to a doubling of CO₂ concentration.

The paper is organized as follows. Section 2 briefly describes the model and experiment design, and the procedure to apply the CFRAM to the NCAR CCSM3-SOM simulations. Section 3 characterizes the El Niño–like response of the model in the equatorial Pacific. The contributions of each feedback process to the equatorial Pacific SST changes are examined in the section 4. A summary and conclusions are then given in section 5.

2. Model setup and analysis approach

The slab ocean model version of the NCAR CCSM3.0 (Collins et al. 2006a) is used in this study. This configuration allows for a fully interactive surface exchange between atmosphere model and ocean model. Therefore, ocean temperature is a prognostic variable. However, the horizontal ocean heat transport and deep-water heat exchange are specified by an internal ocean mixed layer heat flux \( Q \) flux to simplify the ocean model. The thermodynamic sea ice component of the Community Sea Ice Model (CSIM) is employed in this configuration to calculate the sea ice concentration and thickness, surface albedo, and surface exchange flux. The atmosphere model component, the Community Atmosphere Model, version 3 (CAM3; Collins et al. 2006b), is a global atmospheric general circulation model with T42 spectral truncation (approximately 2.8° × 2.8° latitude/longitude) in the horizontal and 26 levels in the vertical from the surface to 2.917 hPa. The land surface model component
is the Community Land Model (CLM), which uses the same horizontal grids as the atmospheric model.

The data from the same two 50-yr equilibrium simulations as those in Song et al. (2014) are used in this study. The present-day climate simulation (referred to as the 1×CO₂ run) uses the 1990 CO₂ value of 355 ppmv, whereas the perturbation climate simulation (referred to as the 2×CO₂ run) uses the doubled CO₂ concentration of 710 ppmv. Both simulations start from the same initial conditions obtained from the NCAR repository. The spatially varying ocean mixed layer depth is specified from Levitus et al. (1998). The monthly mean distribution of internal ocean mixed layer heat flux (Q flux) is derived from the net ocean surface energy budget of 10 years (1981–90) of Atmospheric Model Intercomparison Project (AMIP)-type simulation of the CAM3. The model attains equilibrium with the doubled CO₂ concentration after 25 years of time integration. Thus, the climate statistics and climate feedback analysis presented in this paper are based on hourly output from years 30 to 39 of the simulations.

The CFRAM analysis is applied to the CCSM3-SOM model output as described in Song et al. (2014) to determine the contributions of CO₂ forcing and individual feedback to the temperature changes. For application details readers are referred to Song et al. (2014). The CFRAM is an offline postprocessing diagnostic tool based on the energy balance in an atmosphere–surface column. Any dynamical, thermodynamic, or radiative process that influences the local energy budget is considered a feedback process. The energy perturbation due to any dynamical or physical process between the 1×CO₂ and 2×CO₂ runs is determined by the difference in the corresponding energy flux convergence, which is calculated by the CCSM3-SOM at each time step during the model integration. The radiative energy perturbation can be further decomposed into perturbations due to the doubling of CO₂, temperature change, water vapor, cloud, and albedo feedbacks, which is accomplished by a series of offline radiation calculations using radiation code and hourly output from CCSM3-SOM with a two-sided partial radiative perturbation method (Song et al. 2014). Since the temperature change only influences the longwave radiation, the partial temperature change due to any individual feedback process is derived by requiring the infrared radiation induced by the temperature change alone to exactly balance the energy perturbation due to the feedback process that is considered:

\[ \Delta T_x = \left( \frac{\partial R}{\partial T} \right)_{T = T_0}^{-1} \Delta Q^x, \]  

where \( \Delta \) stands for the difference between the two equilibrium states; \( T = (T_1, T_2, \ldots, T_{\text{surface}}) \) is the vector of temperature from the surface to the top level of the atmospheric model (TOM); and the superscript \( T \) denotes transpose. Also, \( R = (R_1, R_2, \ldots, R_{\text{surface}}) \) is the vector of vertical divergence of net longwave radiation fluxes. The Planck feedback matrix \( \partial R/\partial T \) is determined by the two-sided PRP calculation with 1-K temperature perturbation from each layer. In addition, \( Q^x = (Q_1^x, Q_2^x, \ldots, Q_{\text{surface}}^x) \) is the vector of energy flux convergence associated with feedback process \( x \). Here the superscript \( x \) is substituted for different feedback processes, including conv for atmospheric convection, cond for condensation, adv for ocean process, shf for surface sensible heat flux, lhf for surface latent heat flux, pbl for boundary layer process, gwd for gravity wave drag, adv for dynamical advection process, CO₂ for the external forcing (doubling of CO₂), wv for water vapor feedback, cld for cloud feedback, and albedo for albedo feedback.

3. El Niño–like warming in the CCSM3-SOM

Figure 1a displays the geographical distributions of annual mean SST in the tropical Pacific from the present-day climate simulation. The SST in the tropical western Pacific is higher than in the tropical eastern Pacific. In the equatorial Pacific, a narrow latitudinal band of cold SSTs (cold tongue) appears in the central and eastern Pacific, while the SST warm pool (SST > 28°C) is located west of 170°W. The SST changes due to a doubling of CO₂ from the CCSM3-SOM simulation output (shadings) and the sum of the partial temperature changes due to external forcing and each feedback process (contours) derived using Eq. (1) are shown in Fig. 1b. The CFRAM analysis successfully reproduces the SST change of the CCSM3-SOM in the tropical Pacific. The mean bias in the sum of the partial SST change in the tropical Pacific is 0.009 K, which is negligible (0.5%) compared to the mean SST change from the CCSM3-SOM simulations (1.69 K). The spatial pattern correlation between the two is 0.97. The high accuracy of the CFRAM analysis enables us to quantify contributions of climate feedbacks to SST warming in the tropical Pacific. The SST response in the equatorial Pacific exhibits significant spatial variation with strong warming up to 2.75 K in the central and eastern Pacific and weak warming (typically <1.75 K) in the western Pacific. This warming pattern reduces the mean east–west SST gradient in the equatorial Pacific. Figure 1c further shows the longitudinal distribution of SST change from CCSM3-SOM simulation and CFRAM analysis over the equatorial Pacific (5°S–5°N). It is clear that the CFRAM analysis successfully reproduces the SST warming pattern of CCSM3-SOM in the equatorial...
The SST warming in the equatorial Pacific has a minimum in the western Pacific warm pool between 160°E and 170°W, and gradually increases eastward to its maximum in eastern Pacific between 80° and 100°W. The SST warming difference between the Niño-3 region (5°S–5°N, 90°–150°W; 2.08 K) and equatorial western Pacific (5°S–5°N, 140°E–170°W; 1.57 K) is about 0.51 K, with the largest warming gradient in the central Pacific. Since one of the widely used El Niño definitions is that the SST anomalies of the Niño-3 region are +0.5 K or more for at least six consecutive months (Trenberth 1997), this suggests that the annual mean SST response to a doubling of CO₂ in the slab ocean model version of CCSM3.0 resembles the reduction of SST gradient across the equatorial Pacific during the El Niño event, indicating that the Pacific SST has a tendency toward a more El Niño–like condition as the climate warms.

Associated with the SST changes, the atmospheric convection (precipitation) and circulation also change correspondingly. In the control climate, the warm SST drives deep convection and strong rising motion in the western Pacific (Figs. 2a and 3a). The associated divergent flow in the upper troposphere transports air eastward to central and eastern Pacific in the equatorial plane, which descends to produce a dry and high pressure region over the eastern Pacific cold tongue. The sea level pressure gradient between the high pressure over eastern Pacific and the low pressure over western Pacific...
drives the low-level easterlies, closing the Walker circulation loop (Figs. 2a,c). When the climate warms, the warm SST extends eastward and convection shifts to the central and eastern Pacific, resulting in significant precipitation increase in the central Pacific and the coastal region of eastern Pacific (Fig. 3b). Correspondingly the anomalous ascending motion occurs over the central and eastern Pacific and anomalous descending motion appears over the western Pacific Maritime Continent (Fig. 2b). Meanwhile, the sea level pressure is increased over the central and western Pacific and decreased over the eastern Pacific (Fig. 2d). The weakened sea level pressure gradient between the eastern and western Pacific results in a reduction of low-level easterlies (Fig. 2b). The equatorial westerly in the upper troposphere over central and eastern Pacific is also decreased. All these changes in the equatorial atmospheric circulation tend to oppose the background Walker circulation, slowing down the Walker circulation in the warmer climate. Since the weakening of easterly trade winds and the Walker circulation, the eastward shift of convection over the central equatorial Pacific, and the reduction in SST and SLP gradients across the equatorial Pacific are all major characteristics of El Niño, it suggests that the tropical circulation response resembles an El Niño in the CCSM3-SOM.

4. Contribution of climate feedbacks

The partial SST change due to cloud feedback averaged over the 5°S–5°N latitude belt is shown in Fig. 4a. The net cloud feedback produces SST cooling up to 3.5 K west of 105°W, and SST warming up to 5 K east of 105°W, indicating a negative net cloud feedback over the western and central Pacific and positive cloud feedback over the eastern Pacific. The negative net cloud feedback produces a stronger SST cooling in the central equatorial Pacific than the western Pacific; however, the SST change in CCSM3-SOM shows a greater SST warming in the central Pacific than the western Pacific. This indicates that the cloud feedback tends to damp the El Niño–like warming in the central Pacific and only contributes to the El Niño–like warming in the eastern Pacific east of 105°W. The cloud feedback can be further decomposed into shortwave cloud (cloud albedo) feedback and longwave cloud (cloud greenhouse) feedback. The shortwave cloud feedback shows the same pattern as that of net cloud feedback but with a slightly stronger magnitude, indicating that the cloud feedback is dominated by shortwave feedback and that the longwave feedback tends to damp the impact of shortwave cloud feedback. The shortwave cloud feedback produces SST
warming up to 6 K in the eastern Pacific east of 105°W. It can be largely attributed to the decrease in low-level stratus cloud fraction (Fig. 4b) as a result of increased deep convection in a warmer climate, which allows more solar radiation into the surface. Although the increase of high-level cloud fraction (Fig. 4d) associated with enhanced deep convection may block more incoming solar radiation, the total cloud liquid water path is decreased markedly (Fig. 4c) because of the decrease in low-level cloud, leading to more solar radiation reaching the surface. Both regional-mean cloud liquid water path and high-level cloud fraction are increased in the western and central equatorial Pacific because of more active convection in a warmer climate. However, the increases are larger over the central Pacific than western Pacific resulting from the eastward shift of convection, which blocks more incoming solar radiation and hence results in stronger SST cooling in the central Pacific than western Pacific. The SST change due to longwave cloud feedback is much smaller than that induced by shortwave cloud feedback. It is generally positive in the western and central Pacific but becomes negative in the eastern Pacific east of 105°W, indicating a reduced contribution from longwave cloud feedback to the El Niño–like warming in the eastern Pacific. Clouds absorb upward longwave radiation from the surface and re-emit it at their local temperature. Although clouds at any height emit downward longwave radiation and warm the surface, low-level clouds are more efficient because of their closer proximity to the surface and higher temperature than high-level clouds. In the western and central Pacific, the low-level cloud fraction is increased (Fig. 4b) so that it can absorb and re-emit more longwave radiation to the surface, resulting in a positive longwave cloud feedback and surface warming. In the eastern Pacific east of 105°W, the decrease of low-level cloud leads to less downward longwave radiation onto the surface, resulting in a negative longwave cloud feedback and surface cooling.

Water vapor is the dominant greenhouse gas. The positive water vapor feedback is the largest contributor to global mean surface temperature increase in response to a doubling of CO₂ in the CCSM3-SOM (Song et al. 2014). Over the equatorial Pacific, the water vapor feedback still plays a pivotal role in determining the regional mean SST response. The mean SST change over the equatorial Pacific due to water vapor feedback is as high as 4.3 K, owing to the abundant water vapor supply in the tropics. The water vapor feedback (Fig. 5a) also shows distinct longitudinal variation. The stronger positive water vapor feedback produces larger SST warming in the central and eastern equatorial Pacific than the western Pacific, indicating a positive contribution to the El Niño–like warming pattern. The strength of the water vapor feedback depends on changes of moisture content in the atmosphere. Figure 5b shows the
Fig. 4. (a) Partial SST change (K) due to cloud feedback, (b) change in low-level cloud fraction (%), (c) change in cloud liquid water path (g m$^{-2}$), and (d) change in high-level cloud fraction (%) from 1 × CO$_2$ to 2 × CO$_2$ conditions averaged between 5°S and 5°N.
longitude–pressure cross section of specific humidity changes averaged between 5°S and 5°N. It is clear that the model simulates more increase in moisture over the central and eastern Pacific than the western Pacific. The stronger moistening of middle troposphere over the central and eastern Pacific is mainly as a result of the eastward shift of convection (Fig. 3b) and associated weakening of the Walker circulation (Fig. 2b). The enhanced convection and anomalous rising motion of the Walker circulation over the central and eastern Pacific transport more moisture upward. In the lower troposphere, the weakened easterly (Fig. 2b, shadings) transports less moisture from the central Pacific to the western Pacific, resulting in more increase of moisture over the central Pacific than the western Pacific. Over the eastern Pacific east of 100°W, the change of wind speed in the lower troposphere is relatively small compared to the central Pacific. However, the much stronger ocean evaporation (Fig. 6b) associated with more intense SST warming (Fig. 1b) still supplies more moisture in the near-surface atmosphere. In short, the changes in convection, Walker circulation, and surface evaporation in response to a doubling of CO₂ produce stronger moistening of the troposphere over the central and eastern equatorial Pacific than over the western Pacific. The associated water vapor feedback results in greater SST warming there, leading to an El Niño–like SST response. Similar changes in moisture and water vapor feedback are also observed in the 1987 ENSO warming from La Niña to El Niño conditions (Soden 1997). Note that in Fig. 4b the low-level cloud is decreased in the eastern Pacific east of 105°W even though specific humidity is increased in the lower troposphere. In the CCSM3-SOM, the convective and layered cloud fractions depend on convective mass flux and relative humidity, respectively. The marine stratocumulus clouds are diagnosed using an empirical relationship between marine stratus cloud fraction and lower tropospheric static stability defined by the difference in potential temperature between 700 hPa and the surface. We found that the convective cloud fraction is increased below 700 hPa over the eastern Pacific, corresponding to the enhanced convection (Fig. 3b). The lower tropospheric static stability is increased slightly but the relatively humidity is decreased by up to 3% in the lower troposphere over eastern Pacific, indicating that the decrease in low-level cloud fraction over the eastern Pacific is caused by the reduction of relative humidity.

The ocean surface latent heat flux affects SST in two ways. First, it cools the ocean surface as a result of energy loss from the ocean through evaporation. Second, it heats the ocean surface through enhanced downward infrared radiation because the atmosphere becomes warmer by taking up energy from the ocean surface. The net effect of latent heat flux in the CFRAM analysis is thus smaller than that in conventional surface energy budget analyses that only consider evaporative cooling. The partial SST change due to latent heat flux averaged over the 5°S–5°N latitude belt is shown in Fig. 6a. It is generally positive in the central Pacific and negative in the western and eastern Pacific, meaning that the change in latent heat flux amplifies the SST warming in the central Pacific but damps it in the western Pacific. Therefore, the latent heat flux can contribute to the El Niño–like SST warming in the central and western Pacific.

Figure 6b shows the surface latent heat flux changes. There is a decrease of latent heat flux (evaporation) from the sea surface in the central equatorial Pacific, and an increase in the western and eastern Pacific. Because evaporation depends on surface wind speed, SST, and near-surface humidity, we can qualitatively understand the pattern of latent heat flux change by analyzing the changes of these factors. The SST warming along the equator exhibits a maximum in the eastern Pacific and a minimum in the western Pacific. Consequently the model shows a larger enhancement of evaporation cooling in the eastern Pacific than the western Pacific. In the central Pacific where the surface easterly is reduced the most, the evaporation is weakened mainly because of
lower wind speed, although SST is increased moderately. Note that the central Pacific is also the region where convection is increased the most. This negative relationship between convection and surface evaporation is also reported in the observational study of Zhang et al. (1995), which shows that increased convection weakens the surface wind through interaction with large-scale circulation, leading to a decrease in evaporation.

Although the nonradiative atmospheric processes do not directly affect the surface energy budget and hence SST, they can influence the atmospheric temperature change. The associated changes in downward infrared radiation to the surface may affect SST change. The total effects of nonradiative atmospheric processes represent the lapse rate feedback in the traditional radiative feedback analysis. Therefore both surface and atmospheric partial temperature changes due to nonradiative atmospheric processes are analyzed to understand their roles in the El Niño–like SST warming. Figure 7 presents the partial temperature change resulting from convection averaged over 5°S–5°N in the atmosphere and at the surface. There is large warming in the upper troposphere and cooling in the lower troposphere. Convection transports moist static energy upward, producing a net convergence in the upper troposphere and a net divergence in the lower troposphere. Thus, the enhanced convection in the warmer climate results in a warming perturbation in the upper troposphere and cooling perturbation in the lower troposphere, respectively. The longitudinal distribution of partial temperature change is consistent with that of precipitation (convection) change, showing large decreases in temperature lapse rate in the central Pacific between 120° and 170°W and eastern Pacific east of 105°W. The decrease of temperature lapse rate means that more infrared radiation is emitted into the space and less downward infrared radiation at the surface; therefore the enhanced convection tends to cool the surface. The cooling effect of convection on SST is proportional to the magnitude of convection strengthening under 2×CO₂ conditions. The more enhanced convection leads to stronger SST cooling in the central and eastern Pacific than in the western Pacific. Thus, it tends to damp the El Niño–like warming.

The atmospheric dynamical process tends to damp the energy perturbation induced by convection through changing dynamical advection. Figure 8a shows the sum of changes in adiabatic cooling (aω), and vertical advection of sensible energy (−cₚω∂T/∂p) and latent energy (−Lω∂q/∂p). Here ω denotes the grid mean vertical velocity in pressure coordinate p, a is specific volume
The enhanced upward motion (Fig. 2b) and vertical gradient of latent energy (as indicated by moisture change in Fig. 5b) result in a positive energy advection in the equatorial troposphere (not shown). Meanwhile the adiabatic cooling term, which increases with height, is also strengthened as a result of enhanced upward motion. The net effect of changes in adiabatic cooling and vertical advection of sensible and latent energy is positive energy perturbation in the lower troposphere and negative perturbation in the upper troposphere. It generally explains the pattern of temperature change associated to dynamical advection in the equatorial troposphere, namely warming in the lower troposphere and cooling in the upper troposphere. It generally explains the pattern of temperature change associated to dynamical advection in equatorial troposphere, namely warming in the lower troposphere and cooling in the upper troposphere. The consequent increase of temperature lapse rate leads to less emission of infrared radiation to the space and more downward infrared radiation at the surface, and thereby temperature warming at the surface. The longitudinal distribution of the partial temperature change associated with dynamical advection shows strong SST warming in the central Pacific and relatively weak warming in the western and eastern Pacific (Fig. 8c). This suggests that the dynamical advection contributes to the El Niño–like warming in the central Pacific.

The vertical mixing in the planetary boundary layer (PBL) acts to transport energy from the surface upward to the upper PBL. This would heat and moisten the upper PBL and cool and dry the near surface layer. The PBL is influenced by both upward energy flux from the surface below and atmospheric convection. The surface flux provides energy to the PBL and tends to deepen the PBL with stronger surface heating. The deep convective updrafts act to vent mass/energy out of the PBL, and the convective downdrafts act to dry and cool the PBL. In addition, compensating subsidence around convection in the free troposphere acts to dry the air above the PBL height. Therefore, the stronger convection tends to reduce the height of the PBL. The PBL change in a warmer climate can be approximately measured by the PBL height. The mean change of PBL height
between 5°S and 5°N (Fig. 9a, solid) shows a decrease in the western to central Pacific, with maximum reduction in the central Pacific, and an increase in the eastern Pacific east of 95°W. It generally correlates with the change of convective precipitation (Fig. 9a, dotted) negatively in the western and central Pacific, with the maximum decrease of the PBL height corresponding to the maximum increase of convective precipitation in the central Pacific.
In the eastern Pacific east of 95°W, the convective precipitation is increased. However, the PBL height is not decreased correspondingly. The increase of PBL height can be largely attributed to the strong increase of surface latent heat flux (Fig. 6b). The reduced vertical mixing in the PBL transports less energy upward, resulting in cooling perturbation in the upper PBL and heating perturbation in the layer near the surface in a warmer climate (Fig. 9b). Consequently, the increase in temperature lapse rate leads to less emission of infrared radiation upward and more downward infrared radiation at the surface, and thereby SST warming. In contrast, the enhanced PBL vertical mixing over the eastern Pacific transports more energy upward, leading to anomalous heating in the upper PBL and anomalous cooling in the layer near the surface. The consequent decrease in temperature lapse rate results in SST cooling. Since the largest weakening of PBL process occurs in the central Pacific, there is a stronger SST...
warming in the central Pacific than the western Pacific (Fig. 9c). This suggests that the change in PBL processes contributes to the El Niño–like warming in the central Pacific. Because the cloud base of most of tropical oceanic convection is located in the layer near the surface, the anomalous heating of the layer near surface in a warmer climate over the central Pacific explains the enhanced convection there.

The contributions to El Niño–like SST warming from other processes, such as albedo feedback, large-scale condensation, sensible heat flux, gravity wave drag, CO₂ forcing, and ocean processes, are relatively small and will not be discussed in detail here. To quantify the contribution of each feedback process to the El Niño–like SST warming in the equatorial Pacific, the regionally averaged partial SST changes due to each feedback process in the western Pacific (140°E–170°W), central Pacific (170°–110°W), and eastern Pacific (110°–80°W) and their differences are shown in Fig. 10. The corresponding values of warming/cooling contribution from each feedback are provided in Table 1. The major players in the equatorial Pacific warming are CO₂ forcing, cloud radiative feedback, water vapor feedback, convection, dynamical advection, PBL vertical diffusion, and surface latent heat flux. However, their relative importance in the western, central, and eastern Pacific varies from region to region. Relative to the western Pacific, changes in dynamical advection produce the largest positive SST warming (1.46 K) in the central Pacific. The warming from PBL process and water vapor feedback also has significant contributions (0.86 and 0.78 K respectively). Latent heat flux further amplifies the SST warming difference by 0.32 K. These four processes together account for a warming difference of 3.4 K between the central and western Pacific. Convection and cloud feedback cool the central Pacific by −1.48 and −1.41 K respectively, relative to the western Pacific. For the eastern Pacific, the single most significant contributor to the warming difference from the western Pacific is cloud feedback at 3.76 K followed by water vapor feedback at 0.75 K. The negative contributors are convection and PBL processes. Note that PBL turbulence mixing contributes positively to warming in the central Pacific, but negatively in the eastern Pacific. In contrast, cloud feedback contributes negatively in the central Pacific, but positively in the eastern Pacific. Dynamical advection has negligible contribution to the warming difference between the eastern and western Pacific whereas it has a large positive contribution in the central Pacific. These contrasts clearly show that the mechanisms of the El Niño–like SST warming are different between

![Fig. 10. Regionally averaged partial SST change due to each feedback process between 5°S and 5°N in the western Pacific (140°E–170°W), central Pacific (170°–110°W), and eastern Pacific (110°–80°W) and their differences (K).](image)

<table>
<thead>
<tr>
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<tr>
<td>CO₂</td>
<td>0.81</td>
<td>0.88</td>
<td>0.80</td>
<td>0.07</td>
<td>−0.01</td>
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<tr>
<td>Cloud</td>
<td>−1.29</td>
<td>−2.70</td>
<td>2.47</td>
<td>−1.41</td>
<td>3.76</td>
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<tr>
<td>Water vapor</td>
<td>3.84</td>
<td>4.62</td>
<td>4.59</td>
<td>0.78</td>
<td>0.75</td>
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<tr>
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<td>−5.67</td>
<td>−6.59</td>
<td>−1.48</td>
<td>−2.4</td>
</tr>
<tr>
<td>PBL</td>
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<td>1.43</td>
<td>−0.65</td>
<td>0.86</td>
<td>−1.22</td>
</tr>
<tr>
<td>Dynamical advection</td>
<td>1.97</td>
<td>3.43</td>
<td>1.90</td>
<td>1.46</td>
<td>−0.07</td>
</tr>
<tr>
<td>Latent heat flux</td>
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<td>0.07</td>
<td>−0.27</td>
<td>0.32</td>
<td>−0.02</td>
</tr>
<tr>
<td>Sensible heat flux</td>
<td>0.035</td>
<td>−0.002</td>
<td>−0.002</td>
<td>−0.037</td>
<td>−0.037</td>
</tr>
<tr>
<td>Ocean process</td>
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<td>−0.14</td>
<td>0.06</td>
<td>−0.22</td>
<td>−0.02</td>
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<tr>
<td>Net change</td>
<td>1.57</td>
<td>1.92</td>
<td>2.30</td>
<td>0.35</td>
<td>0.73</td>
</tr>
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</table>
the central and eastern Pacific. To summarize, the changes in dynamical advection is the largest contributor to the El Niño–like SST warming in the central Pacific. The changes in PBL diffusion and water vapor feedback further enhance the warming gradient, with some contributions from surface evaporation change. In the eastern Pacific, cloud feedback is the dominant contributor, with additional contribution from water vapor feedback.

5. Concluding remarks

The simulation of the slab ocean version of CCSM3.0 in response to a doubling of CO$_2$ exhibits an El Niño–like change in the mean state of the equatorial Pacific. The SST warming in the central and eastern Pacific is higher than in the western Pacific, with the largest warming gradient in the central Pacific. Associated with the SST changes, convection shifts to the central and eastern Pacific, resulting in significant precipitation increase there. Correspondingly the Walker circulation is weakened markedly. The anomalous ascending motion occurs over the central and eastern Pacific and anomalous descending motion appears over the western Pacific. The low-level easterlies are reduced as a result of the weakened sea level pressure gradient between the western and eastern Pacific.

The CFRAM is applied to 10 years of hourly output of NCAR CCSM3-SOM to identify factors contributing to the El Niño–like warming in the equatorial Pacific in response to a doubling of CO$_2$. Unlike the traditional surface heat budget analysis, which cannot isolate the contributions of each radiative climate feedback and physical and dynamical process of GCMs, the CFRAM considers every physical and dynamical process that responds to temperature change and affects Earth's energy budget as a climate feedback process, and is able to explicitly quantify the contributions of each feedback to temperature changes. In the tropical Pacific, the mean bias in the sum of partial SST changes due to each feedback derived with CFRAM is negligible (0.5%) compared to the mean SST change from the CCSM3-SOM simulations. The spatial pattern correlation between the two is as high as 0.97.

The analysis shows that the mechanism of El Niño–like SST warming in the central Pacific is different from that in the eastern Pacific. The dynamical advection feedback is the largest single contributor to El Niño–like warming in the central Pacific. The net effect of enhanced adiabatic cooling and vertical advection of sensible and latent energy in the central Pacific results in more increase in temperature lapse rate in the central Pacific, leading to greater SST warming in the central Pacific than in the western Pacific. The maximum weakening of PBL vertical mixing occurs in the central Pacific, making it the second largest contributor to El Niño–like SST warming in the central Pacific. The stronger moistening related to more enhanced convection, stronger anomalous rising motion, and weakened easterlies result in greater positive water vapor feedback in the central Pacific than in the western Pacific. Thus, water vapor feedback also makes a significant contribution to the El Niño–like warming in the central Pacific. The reduced evaporation in the central Pacific due to weakened easterlies leads to warming in the central Pacific and enhanced evaporation as a result of increased SST leads to cooling in the western Pacific, thus also contributing to the El Niño–like warming in the central Pacific. In contrast, the dominant contributor to El Niño–like SST warming in the eastern Pacific is cloud feedback. Both positive cloud feedback associated with low-level cloud decrease in the eastern Pacific and negative cloud feedback due to the increase of cloud in the western Pacific contribute to the El Niño–like warming. In addition, the stronger enhancement of evaporation and convection in the eastern Pacific results in more increase in moisture than the western Pacific, making water vapor feedback also important to the El Niño–like warming in the eastern Pacific.

How does the equatorial Pacific warming influence the precipitation pattern? It is apparent that the pattern of precipitation changes is significantly different from that of SST changes. The maximum equatorial SST warming occurs in the eastern Pacific east of 100°W between 5°S and 5°N (Fig. 1b), while the significant increase of precipitation occurs in the central Pacific and the eastern Pacific within 3°–5°N, 80°–90°W (Fig. 3b). It should be pointed out that the change of convection depends not only on SST change but also on the value of SST in a warmer climate. Studies have shown that a necessary condition for the development and maintenance of strong deep convection in the tropics is that the local SST be about 28°C or greater (Graham and Barnett 1987; Clarke 2008). This high SST ensures that the cloud-base air mass will be charged with the required moist static energy so that convection can reach the upper troposphere (Sud et al. 1999). The SST in the eastern Pacific cold tongue between 5°S and 2°N is generally between 22° and 26°C in the 1×CO$_2$ simulation, whereas it is greater than 26.5°C in the central Pacific west of 150°W and eastern Pacific within 3°–5°N, 80°–90°W (Fig. 1a). Thus even with maximum warming, the SST in the eastern Pacific cold tongue is still below 28°C, which is much lower than the SST in the central Pacific west of 150°W and eastern Pacific within 3°–5°N, 80°–90°W. This qualitatively explains why a significant increase in precipitation only occurs in the central
Pacific and northern equatorial eastern Pacific rather than in the region with maximum warming. To quantitatively represent the impact of warming pattern on precipitation change, Fig. 11a shows the distribution of annual mean changes of convective available potential energy (CAPE), which is used as the closure for the convection scheme (Zhang and McFarlane 1995; Zhang et al. 1998) to determine the intensity of convection. CAPE is calculated as the vertical integral of parcel’s buoyancy from the level of free convection to the equilibrium level, assuming that the parcel is lifted from the lowest model level (~992 hPa). It is increased everywhere with tropical warming. The pattern of CAPE change correlates well with that of precipitation with a correlation coefficient of 0.68, indicating that the precipitation change in the equatorial Pacific can be generally determined by CAPE change. However, the precipitation decrease over the western Pacific Maritime Continent (Fig. 2b) cannot be attributed to CAPE change, indicating that other factors may also influence the precipitation pattern change. One such factor is the low-level moisture convergence. Figure 11b shows the annual mean change of low-level moisture convergence. The anomalous moisture convergence occurs in the central Pacific and northern equatorial eastern Pacific, contributing to the precipitation increase, while the anomalous moisture divergence over northern western Pacific west of 140°W results in the decrease of precipitation.

In this study the ocean dynamics are excluded from the slab ocean model version of CCSM3.0. How will the ocean dynamics influence the warming pattern and climate feedbacks in the fully dynamical ocean model version of CCSM3.0 (CCSM3)? It should be noted that the atmospheric feedback mechanisms (e.g., convection–SST interaction) identified by the CFRAM in the CCSM3-SOM represent the response and feedback of atmospheric processes to temperature change, which are intrinsic characteristics of atmosphere model and independent of the ocean dynamics. So if atmospheric processes are the main drivers of the El Niño–like warming—in other words, if the warming pattern is not substantially changed in the CCSM3 relative to the CCSM3-SOM—the contribution of atmospheric feedbacks in the CCSM3 should be similar to that derived from the CCSM3-SOM. Nevertheless, the ocean dynamical advection may help shape the SST warming pattern. It is shown that the annual mean surface air temperature changes between 1980–99 and 2080–99 in the A1B scenario from the CCSM3 also display an El Niño–like warming pattern (Meehl et al. 2006), with greater warming in the Niño-3.4 region than the equatorial western Pacific (Fig. 5a of Meehl et al. 2006). However, the warming in the far eastern Pacific (east of 100°W) is comparable to that of the western Pacific. Since the maximum warming occurs in the equatorial far eastern Pacific in the CCSM3-SOM, it suggests that

![Figure 11](image-url)
ocean dynamics mainly influence the El Niño–like warming in the far eastern Pacific and acts to damp it. Therefore, we expect the conclusions derived from the CCSM3-SOM to be generally applicable to the fully coupled CCSM3 in the western and central Pacific. In the far eastern Pacific, low-level stratus clouds are prevalent over cold SST tongue and the middle troposphere is very dry. In contrast, deep convective clouds are dominant over the western Pacific warm SST so that the middle troposphere is moist. Therefore, even though the warming amplitude is comparable between the far eastern Pacific and the western Pacific in the CCSM3, the decrease of low-level stratus clouds and increase of moisture in the middle troposphere in the far eastern Pacific can still be more effective than the western Pacific. This indicates that cloud feedback and water vapor feedback, the major contributors to the El Niño–like warming in the eastern Pacific in the CCSM3-SOM, can still produce stronger warming in the far eastern Pacific in the CCSM3 as well. On the other hand, a similar initial SST warming perturbation associated with a doubling of CO₂ may lead to much greater increase in vertical ocean temperature gradient in the far eastern Pacific, where the thermocline is shallow, than in the western Pacific. This may in turn result in much stronger cold water upwelling and surface cooling in the far eastern Pacific in the fully coupled CCSM3. In this way the ocean dynamical process may damp the tendency of enhancing the east–west SST gradient caused by the cloud and water vapor feedback even if the Walker circulation is weakened, resulting in a comparable warming between the far eastern Pacific and the western Pacific in the fully coupled CCSM3.

A recent study (Yeh et al. 2012) shows that the patterns of tropical Pacific SST warming trend over the second half of the twentieth century have changed from a La Niña–like structure in the CMIP3 multimodel ensemble dataset to an El Niño–like structure in the phase 5 of CMIP (CMIP5) dataset. Applying the CFRAM climate feedback analysis to both CMIP3 and CMIP5 models should be able to help us identify which changes in the CMIP5 model result in such change in the SST warming pattern in the tropical Pacific, and improve our understanding of the model uncertainties.

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REFERENCES


