Asymmetric Response of the Equatorial Pacific SST to Climate Warming and Cooling

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ABSTRACT

The response of the equatorial Pacific Ocean to heat fluxes of equal amplitude but opposite sign is investigated using the Community Earth System Model (CESM). Results show a strong asymmetry in SST changes. In the eastern equatorial Pacific (EEP), the warming responding to the positive forcing exceeds the cooling response to the negative forcing, whereas in the western equatorial Pacific (WEP) it is the other way around and the cooling surpasses the warming. This leads to a zonal dipole asymmetric structure, with positive values in the east and negative values in the west. A surface heat budget analysis suggests that the SST asymmetry mainly results from the oceanic horizontal advection and vertical entrainment, with both of their linear and nonlinear components playing a role. For the linear component, its change appears to be more significant over the EEP (WEP) in the positive (negative) forcing scenario, favoring the seesaw pattern of the SST asymmetry. For the nonlinear component, its change acts to warm (cool) the EEP (WEP) in both scenarios, also favorable for the development of the SST asymmetry. Additional experiments with a slab ocean confirm the dominant role of ocean dynamical processes for this SST asymmetry. The net surface heat flux, in contrast, works to reduce the SST asymmetry through its shortwave radiation and latent heat flux components, with the former being related to the nonlinear relationship between SST and convection, and the latter being attributable to Newtonian damping and air-sea stability effects. The suppressing effect of shortwave radiation on SST asymmetry is further verified by partially coupled overriding experiments.

1. Introduction

A majority of climate models tend to project that the mean climate condition in the tropical Pacific will shift toward an El Niño–like state in a warmer climate (e.g., Knutson and Manabe 1995; Liu et al. 2005; Held and Soden 2006; Zhang and Li 2014). This shift of the mean climate state could alter the amplitude, frequency, seasonal cycle, and spatial patterns of ENSO and then affect a range of weather phenomena (e.g., Collins et al. 2010; Vecchi and Wittenberg 2010; DiNezio et al. 2012; Santos et al. 2013; Cai et al. 2015; Zheng et al. 2016).

It is well known that climate is a highly nonlinear system, and fully identifying the nonlinear response of the climate system can be complicated (Yang and Wang 2008). Extensive studies have discussed the nonlinear response in the climate, with a focus on ENSO strength and duration (e.g., Hoerling et al. 1997; An and Jin 2004; An 2009; Kessler 2002; Okumura and Deser 2010; Okumura et al. 2011; Chen et al. 2016). ENSO nonlinearity is usually considered through its asymmetric character; for example, the SST anomalies over the cold tongue region are found to be larger in magnitude during the warm phase (i.e., El Niño) compared to the cold phase (i.e., La Niña), and the center of SST anomalies along the equatorial Pacific is eastward displaced. In
addition to the spatial patterns, the duration of ENSO phases also appears to be asymmetric, with the La Niña events persisting longer than El Niño events. Several hypotheses have been proposed to explain these asymmetric features between El Niño and La Niña. Hoerling et al. (1997) found that nonlinear dependence of atmospheric deep convection on SST may be responsible for the ENSO asymmetry (i.e., the convection anomalies appear over the central and eastern Pacific during El Niño whereas the anomalies associated with La Niña are shifted farther westward). Similarly, Kang and Kug (2002) demonstrated that the relatively weak SST anomalies during La Niña are related to the westward displacement of the wind stress anomalies. Okumura et al. (2011) also found that the nonlinear response of deep convection contributes to the asymmetric evolutions of El Niño and La Niña. Jin et al. (2003) proposed that the ENSO amplitude asymmetry is related to the nonlinear temperature advection by ocean currents, in which the nonlinear term contributes to warming the upper tropical Pacific for both phases of ENSO and thus the strength of El Niño is enhanced but the strength of La Niña is suppressed. In addition, asymmetric thermal heating due to tropical instability waves (TIWs) is also found to have connection with the ENSO asymmetry (An 2008). More specifically, cold ENSO events are usually accompanied by stronger TIWs, which act to ventilate the cold tongue region with warm off-equator temperature, thus producing a damping effect on the cold temperature anomalies. However, this damping effect from the TIWs is weaker during El Niño due to the reduced meridional temperature gradient in the tropical Pacific, leading to the asymmetry between the warm and cold phases of ENSO. Additional processes have also been suggested for the ENSO asymmetry, including the biological-physical feedback (Timmermann and Jin 2002) and the nonlinear response of westerly wind burst to different ENSO phases [Eisenman et al. 2005; see An (2009) for a review].

Furthermore, the climatological mean state over the tropical Pacific appears to contribute to the ENSO asymmetry. For example, An (2009) found that the skewness of ENSO underwent significant interdecadal variations, with negative skewness prevailing before 1960s but positive skewness in more recent decades. According to An and Jin (2004) and Santoso et al. (2013), this change in skewness is closely related to the tendency of eastward propagation in the ENSO variability as climate warms, since the eastward-propagating ENSO tends to produce large nonlinear advective heating and hence intensify the asymmetry of ENSO.

The equatorial Pacific plays a crucial role in modulating the global climate, and the mean SST in the equatorial Pacific will change in response to external forcings and in turn force changes in other components of the climate system (Schneider et al. 1997; Barsugli et al. 2006; Kosaka and Xie 2013, 2016). Therefore, it is important to examine the response of the equatorial Pacific to different external forcings. Recent studies showed that regional oceanic responses to changes in greenhouse gases (GHGs) and aerosols are similar but opposite in sign (Xie et al. 2013). For example, over the tropical Indian Ocean, while GHG induces a positive Indian Ocean dipole-like response, the mean climate condition shifts toward a negative Indian Ocean dipole-like state under aerosol forcing (Li and Luo 2017, manuscript submitted to Atmos.–Ocean). In the North Pacific, while the GHG effect gives rise to a warming of surface temperature and less production of mode waters, the aerosol effect brings about a cooling of surface temperature and more production of mode waters (Wang et al. 2013; Li and Luo 2016). Further, these changes described above are found to be asymmetric, with the aerosol effect exceeding the GHG effect (Li and Luo 2016; Li and Luo 2017, manuscript submitted to Atmos.–Ocean).

By employing NCAR’s Community Earth System Model version 1.1 (CESM1.1), in this study we impose heat fluxes of equal amplitude but opposite sign into the ocean surface to examine the response of the equatorial Pacific to heating versus cooling. As will be shown later, there appear asymmetric changes in many of surface and subsurface fields over the equatorial Pacific Ocean. Our main purpose of this study is to investigate the nonlinearity of the equatorial Pacific mean state in response to external forcings as well as the effects of ocean dynamics and air–sea interaction in modulating the nonlinearity. The rest of the paper is structured as follows. Section 2 describes the experiment design for the fully coupled experiments, the slab ocean experiments, and the partially coupled overriding experiments. Section 3 presents the asymmetric changes over the equatorial Pacific under the warming and cooling scenarios. In section 4, we conduct a mixed-layer heat budget analysis to examine the roles of oceanic horizontal advection, vertical entrainment, diffusion, and air–sea heat flux for the SST asymmetry in the equatorial Pacific, and analyze the results from slab ocean and partially coupled overriding experiments to verify the impact of ocean dynamical processes and shortwave radiation on the SST asymmetry, respectively. The asymmetric changes of the seasonal cycle are also discussed in this section. Finally, a summary and a discussion of our findings are given in section 5.
2. Experiment design

a. Fully coupled experiments

CESM1.1 comprises the Community Atmospheric Model version 5 (CAM5), the Community Land Model version 4 (CLM4) and the Parallel Ocean Program version 2 (POP2). The horizontal resolution of CAM5 and CLM4 is 1.9° longitude × 1.9° latitude, with the atmospheric component having 30 vertical levels. The horizontal resolution of POP2 is nominally 1°, telescoped meridionally to ~0.3° at the equator. Vertically, it has 60 uneven levels with the thickness varying from 10 m near the surface to 250 m at the bottom.

Starting from an equilibrium state that is available at NCAR, a control simulation (CTRL; Table 1) is integrated for 250 years with no external forcing in the coupled atmosphere–ocean system. The heating and cooling simulations are then carried out by adding a uniform heat flux \( Q^* \) of 6 and \( -2 \) \( \text{W} \cdot \text{m}^{-2} \) into the ocean surface, and they are called HEAT and COOL (Table 1), respectively. In such a way, the total surface heat flux into the ocean \( Q_t \) can be expressed as

\[
Q_t = Q_{ao} + Q^*,
\]

where \( Q_{ao} \) is the air–sea surface heat flux, which comprises the longwave radiation \( Q_L \), shortwave radiation \( Q_s \), latent heat flux \( Q_E \), and sensible heat flux \( Q_{SH} \). It should be noted that these heat fluxes are computed interactively every time the atmosphere and ocean communicate through the coupler. The responses to the heating and cooling forcings are taken as HEAT – CTRL and COOL – CTRL, respectively.

Figure 1 shows that the upper ocean reaches a quasi-equilibrium stage after 100 years in both HEAT and COOL and its response to the heating and cooling is nearly symmetric. However, the deep ocean is still adjusting (dashed lines in Fig. 1b), and the adjustment is faster in COOL than HEAT. This is due to the inherent nonlinearity of convection, in which cold, dense surface water will immediately sink while a warm anomaly will float (Xie and Vallis 2012; Manabe et al. 1991).

Although the global-mean SST response is nearly symmetric for the HEAT and COOL experiments

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**TABLE 1. Experiments with CESM1.1.**

<table>
<thead>
<tr>
<th>Name</th>
<th>Run (yr)</th>
<th>Description</th>
</tr>
</thead>
<tbody>
<tr>
<td>CTRL</td>
<td>250</td>
<td>Control run</td>
</tr>
<tr>
<td>HEAT</td>
<td>250</td>
<td>Adding uniform 6 ( \text{W} \cdot \text{m}^{-2} ) to the ocean</td>
</tr>
<tr>
<td>COOL</td>
<td>250</td>
<td>Extracting uniform 6 ( \text{W} \cdot \text{m}^{-2} ) from the ocean</td>
</tr>
<tr>
<td>CTRL_SOM</td>
<td>75</td>
<td>Slab ocean control run</td>
</tr>
<tr>
<td>HEAT_SOM</td>
<td>25</td>
<td>Adding uniform 6 ( \text{W} \cdot \text{m}^{-2} ) to the slab ocean</td>
</tr>
<tr>
<td>COOL_SOM</td>
<td>25</td>
<td>Extracting uniform 6 ( \text{W} \cdot \text{m}^{-2} ) from the slab ocean</td>
</tr>
<tr>
<td>HEAT_SW</td>
<td>50</td>
<td>Same as HEAT, but shortwave radiation is specified to climatology</td>
</tr>
<tr>
<td>COOL_SW</td>
<td>50</td>
<td>Same as COOL, but shortwave radiation is specified to climatology</td>
</tr>
</tbody>
</table>

**Fig. 1.** Time series of global mean (a) SST (solid) and air–sea surface heat flux changes (dashed), and (b) ocean heat content changes (solid for upper 500-m ocean and dashed for deep ocean below 500 m) for (red) HEAT and (blue) COOL experiments. The times series in (a) is smoothed with a 10-yr running mean filter.
(Fig. 1a), a clear asymmetry is found for the regional patterns of the equatorial Pacific SST (comparing the shading in Fig. 2b to Fig. 2a). Results presented below are averaged from the last 100 years of the simulations, and the sum of the anomalies in the two experiments \[(\text{HEAT} - \text{CTRL}) + (\text{COOL} - \text{CTRL})\] is used to measure their asymmetry. For example, a 3°C warm anomaly in HEAT and a 2°C cold anomaly in COOL compared to CTRL over the same region lead to a 1°C asymmetry.

b. Slab ocean experiments

To assess the role of the ocean dynamical processes in generating the SST asymmetry in the equatorial Pacific, a set of experiments (including CTRL_SOM, HEAT_SOM, and COOL_SOM; see Table 1) are designed and conducted using the slab ocean version of CESM1.1, in which active CAM5 and CLM4 are coupled to a slab ocean model. Forced with a repeating annual cycle ocean heat transport derived from the last 100 years’ average of CTRL, the slab ocean control simulation CTRL_SOM quickly reaches a quasi-equilibrium in about 5 years and is further integrated for 70 years. CTRL_SOM reproduces well the global SST pattern in the fully coupled CTRL (not shown). Starting from the 51th year of the CTRL_SOM, the 25-yr-long HEAT_SOM and COOL_SOM are performed by adding a uniform heat flux \(Q_o\) of 6 and \(-6\) W m\(^{-2}\) into the slab ocean, respectively. Results presented in section 4 are averaged over the last 15 years of the simulations.

Since the ocean heat transports in the two perturbation runs of HEAT_SOM and COOL_SOM are prescribed the same way as the CTRL_SOM run, the ocean dynamical contribution to the SST asymmetry can be inferred by comparing the solutions between the slab ocean experiments and the fully coupled experiments.

c. Partially coupled experiments

To isolate the contribution of shortwave radiation to the SST asymmetry in the equatorial Pacific, two partially coupled experiments, HEAT_SW and COOL_SW (Table 1), are conducted using the partially coupled CESM1.1, which is realized through overriding the time series of a certain variable at the air–sea interface obtained from its fully coupled realization to disable the targeted feedback, and this technique has been successfully used to examine the formation processes of the equatorial Pacific SST in response to global warming (Lu and Zhao 2012; Luo et al. 2015, 2017; Liu et al. 2017).

In the two partially coupled experiments, the full time series of shortwave radiative forcing to the ocean is replaced with a seasonally repeating climatological field computed as the last 100 years’ average of CTRL. In this way, a comparison between the fully and partially coupled experiments will identify whether the shortwave radiation is favorable for the development of the SST asymmetry in the equatorial Pacific. Note that the partially coupled experiments are only run for 50 years, as the east-heavy warming pattern in HEAT_SW and
central-heavy cooling pattern in COOL_SW are already clear after 30 years of integration (not shown).

3. Nonlinear response in spatial structure

Figure 2 shows the response patterns in the HEAT (left) and COOL experiments (center) as well as their asymmetries (right). The heating induces an El Niño–like response pattern in the equatorial Pacific, with temperature changes including the largest SST anomalies confined in the east (Fig. 2a) and an overall shoaling of the equatorial thermocline with more shoaling in the west (Fig. 2g). These changes are similar to what happens in the coupled model with increased GHG in the atmosphere (e.g., Luo et al. 2015; Liu et al. 2017). The cooling, however, produces the maximum SST anomalies in the central Pacific (Fig. 2b) and an overall deepening of the equatorial thermocline with more changes also in the west (Fig. 2h). The sum of the SST responses to the heating and cooling results in a positive asymmetry of 0.3°C in the eastern equatorial Pacific (EEP) but a negative asymmetry of −0.4°C in the western equatorial Pacific (WEP; Fig. 2c). The temperature asymmetry appears to be even more significant at the subsurface ocean (Fig. 2i), especially over the western equator where there is a negative asymmetry of about −1.0°C at depth of 100 m and a positive asymmetry of 0.9°C at depth of 200 m. This feature results from the deepening of the thermocline in response to cooling being greater than its shoaling to heating. These above asymmetric responses of temperature in the equatorial Pacific are very similar to those between the warm and cold phases of ENSO (e.g., Hoerling et al. 1997; Okumura et al. 2011; An 2009; Zhang and Sun 2014). Linked to the asymmetric response of SST anomalies, the location of the maximum of the easterly wind anomalies in COOL shifts westward by about 30° longitude in comparison to that of the westerly anomalies in HEAT (cf. Figs. 2e and 2d). This nonlinear response in the zonal wind stress anomalies can in turn cause larger SST anomalies in the EEP (WEP) in the HEAT (COOL) experiment (Kang and Kug 2002).

Significant convective precipitation anomalies are also found with a positive asymmetry in the EEP and a negative asymmetry in the WEP (contours in Fig. 2c). This nonlinear response in atmospheric convective activities is believed to be mainly associated with the different convective feedback regimes, which are dependent on their background SST conditions (Hoerling et al. 1997; Xie 2009). In particular, over the warm pool region (i.e., WEP) where the mean SST is higher, the intensified convection and increased cloudiness induced by warmer SST in HEAT will reflect more insolation back to space and in turn dampen the initial warming, and thus this negative feedback will prevent the SST and convection anomalies from growing too strong (from thick black dot to thick red dot in Fig. 3a), but this negative feedback becomes less effective when the mean state gets cooler (from thick black dot to thick blue dot in Fig. 3a). On the other hand, in the cold tongue region (i.e., EEP) with low cloud stratocumulus decks, large enough positive SST anomalies in HEAT can push the cloud from stratus regime into convective regime and hence produce substantial convective precipitation (from thick black dot to thick red dot in Fig. 3b), whereas cooling the mean SST further in COOL experiment can only push the climate more toward the stratus regime, which is characterized by a tight linear relationship between precipitation and SST (from thick black dot to thick blue dot in Fig. 3b).
However, this asymmetric response in convection acts to oppose the SST asymmetry. For example, the negative convection asymmetry (fewer clouds) in the WEP would allow more shortwave radiation into ocean and thus induce a positive SST asymmetry. This indicates that the SST–radiation–cloud feedback cannot account for the SST asymmetry in the equatorial Pacific, and more detailed analysis is needed to understand this SST asymmetry.

4. Heat budget analysis

To understand the underlying mechanisms that cause the asymmetric SST response in the equatorial Pacific, we perform a heat budget analysis for the variable mixed layer following the equation in Stevenson and Niiler (1983):

\[
T_t = Q_n + Q_x + Q_y + Q_z + Q_{\text{diff}},
\]

where \( T_t \) represents the mixed layer temperature tendency; \( Q_n \) is the net surface heat flux into the mixed layer, where \( Q_n \) is the net surface heat flux at the bottom of mixed layer and \( \rho_0 \) and \( c_p \) are the density and specific heat of seawater, respectively; \( Q_x \) is the zonal advection, \( Q_y \) the meridional advection, and \( Q_z \) the vertical entrainment. Since the vertical diffusion term \( Q_{\text{diff}} \) is not stored as part of the model’s output and it would be extremely inaccurate to obtain it based on the output data, we use a residual term \( Q_r \) to represent it and close the heat budget. The residual term is comprised of ocean heat transport by unresolved subgrid-scale processes as well as submonthly oceanic processes. As such, the sum of last four terms on the right-hand side of Eq. (2) is the total ocean heat transport \( Q_o \) due to horizontal advection, vertical entrainment, and diffusion, which balances the net surface heat flux into the ocean. All the heat budget terms are defined as downward positive and upward negative. Our modeled climatological heat budget results (not shown) are in good agreement with those calculated with high-frequency observational data in Huang et al. (2010). Surprisingly, though derived as a residual term, our \( Q_r \) is found to have a high similarity to their vertical diffusion term, indicating that this residual term is largely determined by vertical diffusion.

To elaborate on the ocean dynamical effect, we further divide the horizontal advection and entrainment terms into linear and nonlinear parts:

\[
\begin{align*}
Q_x &= -\left(\overline{uu}T_x + uT_x\right) - u'T_x, \\
Q_y &= -\left(\overline{vv}T_y + vT_y\right) - v'T_y, \\
Q_z &= -\left(\overline{ww}T_z + wT_z\right) - w'T_z,
\end{align*}
\]

where \( u \) and \( v \) are zonal and meridional velocity averaged over the mixed layer, \( T_x \) and \( T_y \) are zonal and meridional gradients of mixed-layer temperature, \( w \) is the entrainment velocity across the bottom of the mixed layer, and \( T_z \) is the temperature difference between the mixed layer and the entrained water. The overbar and prime denote the climatological mean and anomaly, respectively. The terms within the bracket on the right-hand side represent the linear dynamic terms (LDTs), and the last terms on the right-hand side represent the nonlinear dynamic terms (NDTs).

4a. Ocean heat transport

Figure 4 shows the spatial distribution of the total ocean heat transport, oceanic horizontal advective terms, vertical entrainment, and residual terms in HEAT and COOL, along with their corresponding asymmetries over the equatorial Pacific. It can be seen that the total ocean heat transport, especially its horizontal advective and vertical entrainment components, favors a positive asymmetry in the EEP and a negative asymmetry in the WEP (right panels in Fig. 4).

For the HEAT experiment, both anomalous zonal and meridional advections contribute to warm the upper ocean (Figs. 4a,d), with the former dominating the response in the western equatorial Pacific and the latter in the eastern equatorial Pacific. The anomalous vertical entrainment, however, exerts a strong cooling effect right on the central and eastern equator (Fig. 4g). A decomposition of the horizontal advections and entrainment terms finds that the anomalous warming in the zonal advection over the WEP region results from the combined effect of decrease in both east–west SST gradient and westward zonal surface current (Fig. 5a), and the anomalous warming in the meridional advection along the central and eastern equator is likewise due to a reduction of both the meridional current and temperature gradient (Fig. 5d). The anomalous cooling in the vertical entrainment over the EEP region is a result of the cooling effect due to increased stratification exceeding the warming effect due to decreased upwelling (Fig. 5f).

For the COOL experiment, it is not surprising that the signs of the anomalous horizontal advections and vertical entrainment (Figs. 4b,e,h) are reversed compared to those in HEAT. However, the anomalous patterns of the horizontal advections are displaced westward by a few tens of degrees, and the vertical entrainment anomalies appear to be larger compared to those in HEAT, implying that the advective responses are not linear to the forcings of the same amplitude but opposite sign.

In terms of the asymmetries of the horizontal advections and entrainment in the two experiments, it is found that, for all the three terms, there are positive asymmetry over the EEP and negative asymmetry over the
WEP (Figs. 4c,f,i), resembling the pattern of the SST asymmetry (Fig. 2c). Next we will decompose them into LDTs and NDTs to further understand the dynamic processes for generating the SST asymmetry.

1) LINEAR AND NONLINEAR DYNAMICAL TERMS

Based on Eq. (3), the horizontal advections and vertical entrainment can be decomposed into the LDTs and NDTs, and their decompositions over the WEP and EEP regions are shown in Fig. 5.

For the EEP region (right panels in Fig. 5), we find that all the three terms contribute to the positive ocean heat transport asymmetry, with meridional advection ($0.06\, ^\circ C \, mon^2$) and vertical entrainment ($0.05\, ^\circ C \, mon^2$) being more important. In more details, the asymmetry in the meridional advection is due largely to the LDTs (Fig. 5d), with its two linear terms ($-y T_0$ and $-y^'T_0$) contributing equally. For the vertical entrainment, while both of its linear terms $-w T_0$ and $-w' T_0$ contribute little to the positive asymmetry (Fig. 5f), the nonlinear term $w T_0$ plays a significant role. For the WEP region (left panels in Fig. 5), the negative ocean heat transport asymmetry results from the zonal advection and vertical entrainment, with the former mainly from NDT ($-0.04\, ^\circ C \, mon^2$) and the latter from one of its linear term $-w T_0$ ($-0.07\, ^\circ C \, mon^2$).

A closer inspection of the NDTs in the equatorial Pacific (Fig. 6) indicates that the total NDT in HEAT contributes to warming the central and eastern equatorial Pacific between $170^\circ$ and $100^\circ$W and to cooling the western equatorial Pacific (contours in the left panels of Fig. 6). The warming is clearly dominated by its vertical component (Fig. 6e), resulting from the weakened entrainment velocity ($w' < 0$) associated with the less vigorous easterly wind and more stratified upper ocean ($T_0' > 0$; Fig. 5f) due to greater warming near the surface (Luo et al. 2009). The NDT cooling in the WEP comes from its zonal component (Fig. 6b), resulting from the weakened zonal current ($u' > 0$) and decreased zonal SST gradient ($T_x' > 0$; Fig. 5a).

It is surprising to see that the contribution of the NDTs in COOL is not opposite, but rather appears to be similar to that in HEAT (comparing the contours in the left panels with right panels of Fig. 6), having a cooling over the far west and a warming over the central and east along the equator. Similarly, the cooling mainly results from its zonal component (Fig. 6a), resulting from the weakened zonal current ($u' > 0$) and decreased zonal SST gradient ($T_x' > 0$; Fig. 5a).
and less stratified upper ocean \((T'_z < 0)\) due to greater cooling near the surface. Therefore, the total positive (negative) NDT over the central and eastern (western) equatorial Pacific contributes positively (negatively) to the warming in HEAT but negatively (positively) to the cooling in COOL, favorable for the development of the SST asymmetry over the equatorial Pacific.

2) RESIDUAL TERM

Although the subgrid-scale and submonthly ocean processes cannot be accurately quantified due to the lack of high-frequency data outputs, the residual term is mostly determined by vertical diffusion. In HEAT, the upper ocean stability in the EEP is increased due to the greater warming at surface than subsurface (Fig. 2g). This suppresses the vertical diffusivity through a Richardson number–dependent parameterization (Pacanowski and Philander 1981), and this cold diffusive flux and entrainment in turn lead to an anomalous warm vertical diffusion (Fig. 4j; Liu et al. 2005; Yang et al. 2009). For the COOL experiment, on the contrary, the decreased ocean stability enhances the vertical diffusivity and thus the residual term is negative in the EEP (Fig. 4k).

The diffusion acts to balance the horizontal advection and entrainment along the equator and thus tends to reduce the SST asymmetry there (Fig. 4l). The larger diffusive anomaly in COOL than the HEAT is due to the fact that the cold surface water mixes more easily than the warm surface water (Manabe et al. 1991).

3) SLAB OCEAN EXPERIMENTS

The abovementioned heat budget analysis suggests the dominant role of ocean dynamical processes in generating SST asymmetry. To verify this, we further perform a set of experiments with a motionless slab ocean (section 2b) in which ocean heat transport is prescribed, and thus the ocean dynamical effect can be inferred by comparing its result with that from the fully coupled experiments.

Figure 7 shows the SST anomalies in the slab ocean HEAT_SOM and COOL_SOM experiments, as well as their corresponding asymmetry, respectively. Without the ocean dynamical effect, the characteristics of the
SST anomalies and asymmetry in the slab ocean show fundamental differences from those in the fully coupled models. For the HEAT_SOM, the warming anomaly is quite uniform over the central and eastern equatorial Pacific (Fig. 7a), whereas it features the EEP-confined warming pattern in the HEAT experiment. For the COOL_SOM experiment, the cooling anomaly is more eastward concentrated (Fig. 7b) compared to that in the coupled COOL experiment. These SST anomalies in the slab ocean experiments result in SST asymmetry in the form of a horseshoe pattern (Fig. 7c). This is in sharp contrast to the SST asymmetry in the fully coupled runs, which features a clear dipole pattern with warming in the east and cooling in the west. It should be noted that the slab ocean experiments exhibit a wind stress asymmetry with easterly anomaly in the WEP and westerly anomaly in the EEP (not shown). When coupled to a dynamical ocean, this wind stress asymmetry would favor the formation of SST asymmetry similar to Fig. 3c through ocean dynamical processes such as weakening the vertical entrainment (warming) in the EEP and vice versa (cooling) in the WEP. Therefore, the slab ocean experiments verify that the ocean dynamical processes are essential for the SST asymmetry found in the fully coupled experiments.

b. Surface heat flux

The ocean heat transport discussed above is balanced by the net surface heat flux, implying that the latter should act to dampen the SST asymmetry at the equatorial Pacific. Figure 8 shows the changes in the air–sea surface heat flux ($Q_{an}$) and its three components in the HEAT and COOL experiments as well as their asymmetries. Note that the sensible heat flux is not shown because of its negligible contribution.
For the HEAT experiment, the equatorial ocean features a heat loss to the atmosphere with a maximum around 130 W (contours in the left panels of Fig. 8), resulting from a combination of warming effect in longwave radiation $Q_L$ (Fig. 8a) and cooling effect in both shortwave radiation $Q_s$ (Fig. 8d) and latent heat flux $Q_E$ (Fig. 8g).

The warming effect from the longwave radiation $Q_L$ can be attributed to a combined effect of water vapor, cloud cover, and SST (Wallace and Hobbs 2006). There are mainly two competing effects to determine the sign of the $Q_L$ anomaly: whereas a warmer ocean tends to emit more upward longwave radiation following the Stephan–Boltzman law, the increased water vapor and cloud cover act to trap more infrared radiation and reflect it back into the ocean. For the HEAT experiment, it turns out that the latter wins out, and thus the net longwave radiation $Q_L$ exerts a warming effect on the equatorial ocean.

The cooling effect from the shortwave radiation $Q_s$ is closely related to increased clouds, especially high-level and midlevel clouds (Figs. 9a,d), reflecting more solar radiation back to space. To dissect the role of latent heat flux in the COOL experiment, $Q_E$ is decomposed into four major terms (e.g., Richter and Xie 2008; Xie et al. 2010): Newtonian damping $Q_{EO}$, wind–evaporation–SST feedback $Q_{EW}$, surface stability $Q_{ESt}$, and relative humidity $Q_{ERH}$. Red bars in Fig. 10 represent their changes over the EEP for HEAT, respectively. It is clear that the cooling effect in $Q_E$ is mainly rooted in $Q_{EO}$, reflecting mainly the effectiveness of a warmer mean SST in cooling the surface through evaporation upon an imposed warming. In addition, $Q_{ESt}$ is also found to be a cooling effect, indicating that the simulated SST increases more quickly than the overlying air temperature. The other two components $Q_{EW}$ and $Q_{ERH}$ act to warm the EEP region, with the former being related to the weakened easterlies (Fig. 2d) and the latter resulting from the increased relative humidity (not shown).

For the COOL experiment, the equatorial ocean shows a heat uptake over the west and the southeast equatorial Pacific (contours in the center panels of Fig. 8). The heat uptake over the former region is due mainly to the warming effect from the shortwave radiation $Q_s$ (Fig. 8e), resulting from more solar insolation in response to a decrease of high- and middle-level cloud cover (Figs. 9b,e). The heat uptake over the latter region is mainly attributed to the warming effect from the latent heat flux $Q_E$ (Fig. 8h), which is in turn dominated by a change in Newtonian cooling $Q_{EO}$ (not shown). In addition, the longwave radiation $Q_L$ plays a cooling role over the equatorial ocean in COOL (Fig. 8b).

1) SHORTWAVE RADIATION

Just the opposite to the total advective asymmetry (contours in Fig. 4c), the total surface heat flux asymmetry (contours in the right panels of Fig. 8) is positive in the WEP (7.1 W m$^{-2}$) and negative in the EEP ($-6.1$ W m$^{-2}$) and thus acts to reduce the SST asymmetry there. More specifically, shortwave radiation is the leading source of the positive asymmetry in the WEP (8.9 W m$^{-2}$) and also contributes about half of the
negative asymmetry in the EEP (−2.8 W m⁻²; Fig. 8f). The asymmetric response of shortwave radiation can be understood as follows. Over the WEP where the climatological SST is higher, while the increase in clouds (Figs. 9a,d) in response to a warming anomaly is limited due to the negative SST–radiation–cloud feedback, the decrease in clouds to a cooling anomaly is barely restricted (Figs. 9b,e). This leaves a negative cloud asymmetry (contours in Fig. 9c) and thus a positive shortwave radiation asymmetry in the WEP. Over the EEP where the climatological SST is lower, the change in clouds due to a cooling anomaly is suppressed because the condition there is already cold and dry (center panels in Fig. 9), whereas the shortwave insolation due to a warming anomaly is significantly enhanced as the deep clouds build up (Figs. 9a,d). This results in a net positive cloud asymmetry (contours in Fig. 9c) and thus a negative shortwave radiation asymmetry there.

2) PARTIALLY COUPLED EXPERIMENTS

The above analysis suggests that the shortwave radiation response (i.e., cloud–radiation–SST feedback) to the heating and cooling in the coupled model is a dominant factor in alleviating the SST asymmetry in the equatorial Pacific. To further demonstrate this, we conduct an overriding technique to the CESM1.1 model to eliminate the contribution of shortwave radiation (section 2c). Figure 11 shows the SST changes in the overriding experiments from the fully coupled experiments. It is clear that, without the modulation of shortwave radiation, both the EEP-centered warming in HEAT (Fig. 11a) and the central and western equatorial Pacific-centered cooling in COOL (Fig. 11b) are significantly enhanced, resulting in a much stronger SST asymmetry along the equator (Fig. 11c). This result is consistent with our previous heat budget analysis, verifying the importance of shortwave radiation in damping the SST asymmetry.

3) LATENT HEAT FLUX

In addition to the shortwave radiation, the latent heat flux $Q_e$ is another contributor to the negative $Q_t$ asymmetry in the EEP (Fig. 8i). According to Fig. 10, the $Q_e$ asymmetry (−3.3 W m⁻²) primarily originates from the asymmetric changes in $Q_{eo}$ (−1.1 W m⁻²) and $Q_{edt}$ (−2.0 W m⁻²) between HEAT and COOL. The mechanism leading to the $Q_{eo}$ asymmetry is easy to
understand (i.e., the SST anomaly sets the strength of Newtonian damping, and a larger SST anomaly in HEAT induces a stronger $Q_{EO}$ relative to that in COOL). The underlying mechanism resulting in $Q_{EdT}$ asymmetry seems to be more complicated. For the COOL experiment, the change of $Q_{EdT}$ is close to zero since the air–sea temperature difference remains largely unchanged. The change of $Q_{EdT}$ ($\pm 2.0 \text{ W m}^{-2}$) in HEAT, in contrast, is to cool the surface ocean since the SST gets warmer than its overlying air temperature by about 0.3°C. Further, the difference in surface stability is likely to be related to the nonlinearity in the responses of atmosphere convection in the EEP (contours in Figs. 2a and 2b). The heating induces a positive convection anomaly there, which enhances the ascending motion of heated sea surface air and thus decreases the air–sea temperature difference. However, the cooling is not able to produce significant changes in the atmospheric convection, and thus the heat exchange between the sea surface and higher air remains almost unchanged, and so does the air–sea temperature difference.

c. Seasonal cycle

In addition to the climatological mean pattern, the seasonal cycle of SST at the equatorial Pacific is also found to have asymmetric response to the uniform heating and cooling. Figure 12 shows the seasonal evolution of the equatorial SST as well as its changes in HEAT and COOL. The climatological SST (Fig. 12a) displays a pronounced annual cycle due to the northward shift of the intertropical convergence zone (ITCZ; Mitchell and Wallace 1992), peaking around April with a maximum of 2.1°C and reaching a minimum of −1.9°C around September. The uniform heating produces a weakening in the seasonal cycle, manifested as greater warming (0.5°C) during cold months and less warming (−0.5°C) during warm months (Fig. 12b), which is similar to what happens in CESM1.1 with increasing GHG in the atmosphere (Liu et al. 2017). Such a change in the seasonal cycle of SST corresponds well with the change in zonal wind stress, which exhibits a robust weakening of its seasonal cycle in the warming climate, with its maximum exceeding 0.01 N m$^{-2}$, roughly one-third of its seasonal change (cf. Figs. 12e and 12d). In the COOL experiment, the SST seasonal cycle is enhanced with larger amplitude and delayed phase; that is, the SST anomaly reaches a maximum (0.6°C) around July and minimum (−0.7°C) around December (Fig. 12c), almost a season behind its climatological SST evolution. This phase delay is associated with the zonal wind change; that is, positive wind anomalies induced by the cooling appear to persist into boreal fall over the central Pacific (Fig. 12f).

To examine the processes that are responsible for the asymmetric changes in the seasonal SST, a monthly mixed-layer heat budget analysis is performed in the EEP region where the seasonal change is strongest. For the HEAT experiment (Fig. 13a), the evolution of temperature tendency $T_t$, featuring positive values from March to October but negative values from November to April, clearly follows the evolution of net surface heat flux $Q_n$, and ocean heat transport contributes little to the weakening of the seasonal cycle. For the COOL experiment (Fig. 13b), however, it appears that ocean heat transport drives the enhancement of the seasonal cycle whereas $Q_n$ acts to cool the surface ocean all year round. Taken together (Fig. 13c), the asymmetric $T_t$ due to ocean heat transport and that resulting from net surface heat flux tend to balance each other (e.g., the heating asymmetry induced by ocean heat transport reaches a maximum during boreal spring when the cooling asymmetry resulting from the net heat flux happens to be a maximum). Overall, the evolution of the asymmetric temperature tendency is clearly dominated by ocean dynamical processes.

5. Summary and discussion

There have been considerable studies on ENSO nonlinearity, with a few different hypotheses proposed to explain asymmetric features between El Niño and La Niña. In this work we have studied the asymmetric response of the equatorial Pacific to energy fluxes of equal amplitude but opposite sign into the ocean surface using CESM1.1.

Our model results show that whereas heating produces an EEP-centered warming pattern, a cooling experiment induces a central equatorial Pacific-centered cooling pattern. This results in a seesaw-like asymmetric pattern.
pattern in SST, with positive values in the EEP and negative values in the WEP. Through a mixed layer heat budget analysis, we find that it is the oceanic horizontal advection and vertical entrainment that generate the asymmetric SST pattern, and both their linear and nonlinear components make contributions. For the linear component, its change in COOL is less (more) significant in the EEP (WEP) compared to that in HEAT, favoring the seesaw pattern of SST asymmetry over the equatorial Pacific. For the nonlinear component, its change acts to warm the central and eastern equator and to cool the western equator in both scenarios, also favorable for the development of the SST asymmetry. The dominant role of ocean dynamical
processes in generating the SST asymmetry is further verified with a pair of heating and cooling experiments with a slab ocean, in which the SST changes and their asymmetry are found to be fundamentally different from those in the fully coupled heating and cooling experiments.

The surface heat flux works to reduce the SST asymmetry through its components of shortwave radiation and latent heat flux. While the asymmetric response of the shortwave radiation acts to warm the WEP region but cool the EEP region, that of the latent heat flux operates to provide additional cooling to the EEP. A more significant SST symmetry between the heating and cooling overriding experiments, in which the shortwave radiation is replaced with the CTRL climatology, further confirms the important role of shortwave radiation in alleviating the SST asymmetry in the equatorial Pacific.

Our model projects an decrease (increase) of the seasonal cycle of SST in the equatorial Pacific in a warmer (colder) climate and these changes are also asymmetric (i.e., the seasonal change in COOL features a larger amplitude and a phase delay compared to that in HEAT). This asymmetric change is also dominated by the ocean dynamical processes.

The heat flux perturbation of 6 W m\(^{-2}\) is chosen for this study since it is comparable to the average forcing under climate change scenario RCP 8.5. Just like the situation between El Niño and La Niña in which the asymmetry depends on their amplitude (Dommenget et al. 2013), we speculate that the asymmetric response presented in this study may also be nonlinear and sensitive to the amplitude of the heat flux forcing. However, an examination of the sensitivity of the asymmetry to the forcing magnitude is beyond the scope of the present study but remains one of interest and further study.

We cannot discount the possibility that some of the above results are a peculiarity to the CESM1.1, considering that some climate models project an increased zonal SST gradient under global warming (Vecchi et al. 2008; Li et al. 2016; Zhang and Karnauskas 2017). However, in view of the considerable parallel between the asymmetry in the mean state changes responding to opposite external forcing and that in ENSO’s warm and cold phases, especially with regard to the established nonlinear behaviors of wind stress anomalies (Kang and Kug 2002; Okumura et al. 2011; Chen et al. 2016) and the associated nonlinear advection by ocean currents (Jin et al. 2003; An and Jin 2004; An 2009), some of the confidence in the asymmetry in the natural variability of ENSO simulated by CESM1.1 may be carried over to the asymmetry of the mean state response under opposite external forcings.

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