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Understanding the ocean temperature change in global warming: the tropical Pacific

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ABSTRACT

The response mechanisms of the tropical Pacific Ocean temperature to increased atmospheric CO_2 are investigated in a coupled climate model. Ensemble simulations are performed under both the transient and stable CO_2 forcing. It is found that the dominant mechanism for temperature change differs in different stages of global warming. During the transient stage, the surface heat flux is a major driving factor for the tropical surface warming. During the equilibrium stage, the dominant mechanism to maintain the surface warming is the meridional advection. The heat flux forcing becomes a damping factor instead, particularly for the western tropical Pacific. Different from the surface warming, the subsurface warming results from the oceanic mixings during the entire period of global warming, whereas the advection terms generally play a cooling role, consistent with the slowdown of the shallow meridional overturning circulation. This paper emphasizes the deterministic role of the dynamic adjustment of the ocean circulation in the long-term change of ocean climate.

1. Introduction

There are numerous studies on the response patterns of climate system to the enhanced greenhouse gases (GHGs) forcing (IPCC, 2007). It is very important to identify the climate change patterns because the patterns provide valuable information on the climate sensitivity regions and the practical guidance for narrowing down our attentions in the future climate research. Although the climate responses to global warming are versatile in different coupled climate models, several robust response patterns have been identified in previous studies (Cubasch et al., 2001), such as the polar amplification, the stronger warming over the land than over ocean and the enhanced equatorial response (EER; Liu et al., 2005).

The response of the ocean to global warming has received great attentions because of its critical role in the long-term climate change (e.g. Levitus et al., 2005; Liu et al., 2005; Pierce et al., 2006). Particular interests have been put on the tropical Pacific Ocean. 'A majority of models show a mean El Niño-like response in the tropical Pacific, with the central and eastern equatorial Pacific sea surface temperature (SST) warming more than the western equatorial Pacific', as stated in the earlier IPCC report (Cubasch et al., 2001). However, in both models and observations, it remains very controversial if the SST trend is El Niño-like warming or La Niña-like warming because, first, many models do not exhibit El Niño-like warming (Liu et al., 2005), and second, the observational tropical Pacific SST trends differ substantially among different data sets and for different time periods (e.g. Knutson and Manabe, 1998; Liu et al., 2005). 'The most likely scenario ($p = 0.59$) is for no trend toward either mean El Niño-like or La Niña-like conditions' (Collins and CMIP Modeling Groups, 2005). Nevertheless, there seems to exist a more robust response pattern of tropical Pacific SST to global warming, the EER, which is characterized by a stronger warming near the equator than in the subtropical Pacific (Liu et al., 2005).

Understanding the fundamental mechanisms of climate response is more important than just knowing the response patterns. This has been a great challenge because of limited reliable observations and the limitations of research approaches and models. For example, the mechanisms of the equatorial Pacific SST response are more controversial than the response patterns themselves. For the El Niño-like warming, the weaker warming in the west has been suggested to be regulated by the negative feedback associated with the latent heat cooling (Knutson and Manabe, 1995) or cloud radiation forcing (Meehl and Washington, 1996; Timmermann et al., 1999) on the warm pool SST. For the La Niña-like warming, the eastern Pacific warming is restrained by the strong oceanic upwelling cooling and, in turn, the wind-upwelling dynamic feedback (Clement et al., 1996;

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Cane et al., 1997) or through equatorward oceanic ventilation (Seager and Murtugudde, 1997; Liu, 1998; Yang and Liu, 2005). For the EER warming, the suppression of equatorial oceanic mixing/entrainment due to the increased surface ocean stability plays a critical role (Liu et al., 2005), with foci on the meridional SST gradient, the atmospheric Hadley circulation and the wind-evaporation feedback.

To gain insight into the ocean response mechanism to external forcing, this paper anatomized the ocean temperature equation. Sensitivity experiments are performed by forcing a coupled climate model with an enhanced CO_2 concentration. It is found that the mechanisms for the ocean temperature changes are different during different stages of CO_2 warming. During the transient stage of global warming, the surface heat flux is a major driving factor for the tropical surface warming, whereas during the equilibrium stage, the dominant mechanism to maintain the surface warming is the meridional advection. The heat flux forcing becomes a damping factor, particularly for the western tropical Pacific. Different from the surface warming, the subsurface temperature change is balanced by the warming factor of oceanic mixings and the cooling factor of the advection terms. The latter is consistent with the slowdown of the shallow meridional overturning circulation.

The response patterns and associated mechanisms examined in this paper are appropriate for our model and could be model-dependent. Our results can be viewed as one manifestation of a broad spectrum of possibilities, and thus the study here could enhance our understanding of the real world and provide clues to further identify and understand the possible future climate changes. This paper is structured as follows: Section 2 introduces the coupled models and experiments. Section 3 summarizes the general changes in the Pacific Ocean. Section 4 investigates the mechanisms of the ocean responses through term balance analyses. Section 5 provides conclusions and discussions. The key conclusion in this work is that because of the slow evolution of ocean, the mechanisms controlling the ocean temperature changes are different in different stages of global warming.

2. Model, experiments and approach

The fast ocean–atmosphere model (FOAM) is used in this paper. The FOAM is developed jointly at University of Wisconsin-Madison and the Argonne National Laboratory (Jacob, 1997). The atmosphere model uses the physics of the National Center for Atmospheric Research (NCAR) community Climate Model 3 (CCM3), with a resolution of R15 and 18 vertical levels. The ocean model was developed following the Geophysics Fluid Dynamics Laboratory (GFDL) Modular Ocean Model (MOM) with a resolution of 1.4° latitude × 2.8° longitude × 32 vertical levels. Without flux adjustment, the fully coupled control simulation (CTRL) has been integrated for over 1000 yr, showing no apparent climate drifts. The FOAM has been used extensively to simulate the global climate for the past, present and future. Comprehensive analyses of the model simulation can be found, for example, in Wu et al. (2003) and Yang and Liu (2005).

The FOAM is computationally efficient so that ensemble experiments can be performed. Two sets of experiments are carried out. The first set is a three-member ensemble run. Starting from a present-day control run with the 1990 CO_2 level of 355 ppmv, the model is forced by a sudden doubling of the CO_2 and is integrated for 200 yr. To investigate the maintaining mechanism of ocean temperature changes during the equilibrium period of global warming, monthly outputs of last 50 yr are examined. The second set is a twenty-member ensemble run. The model is integrated with a 1% yr^{-1} increase in CO_2 for 30 yr. The driving mechanism of ocean temperature changes during the transient period is investigated by analysing the linear trend of ocean temperature. In this paper, the transient 1% CO_2 simulation is referred to as CO2T, whereas the equilibrium double CO_2 simulation is referred to as CO2E. These ensemble experiments should reduce the modelling uncertainty. For the convenience of comparison, parallel control run is also performed for each single experiment. In the following, we will focus on the discussions of the ensemble mean difference between the warming experiments and control runs.

To quantify the contributions of different mechanisms to the temperature change, we analyse the heat budget by calculating the terms in temperature equation (see Appendix A). For the CO2E, the term differences between the warming experiments and control runs are analysed (A2). For the CO2T, the differences between the linear trends of two parallel simulations are analysed (A3). These differences are taken as the anomalous heat budget responsible for the CO_2-induced temperature change.

3. Changes in the tropical Pacific mean state

The transient response of the tropical Pacific SST to the increasing CO_2 shows a linear trend of 0.01°C yr^{-1} (dashed black line, Fig. 1). The equilibrium response to a doubling CO_2 is around 1.4°C (Solid black line, Fig. 1), which is in the range given by the fourth assessment of IPCC (2007). The warming pattern in the Pacific shows an EER (Fig. 2a). The SST within 10° equatorward of tropics is warmed about 1.5°C, and shows a La Niña-like warming in the zonal direction, whereas the SST between 10°–25° in both hemispheres is increased less. The responsible mechanisms are believed to be the suppression of equatorial oceanic mixing/entrainment due to the increased surface ocean stability in the tropics (Liu et al., 2005) and the enhanced latent heat loss due to a strong sensitivity of latent heat flux in the subtropics (Seager and Murtugudde, 1997). The largest warming occurs in the high latitudes of the two hemispheres (Fig. 2a). There is a 2.5°C warming in the subpolar region of the Northern Hemisphere, which may result from a sea-ice albedo positive feedback.

The subsurface temperature change is different from the surface in both the magnitude and pattern. The transient

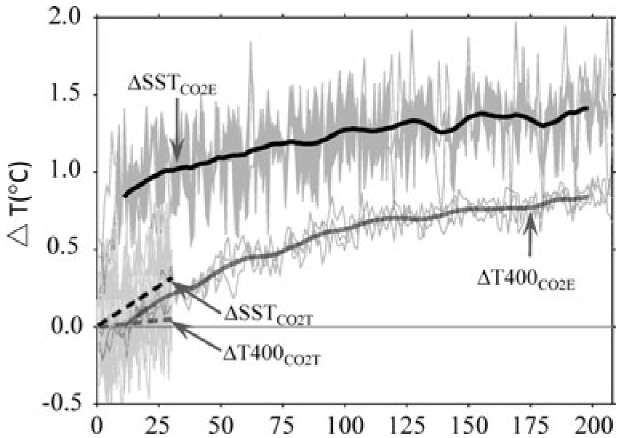

Fig. 1. Evolution of anomalous annual mean SST (black lines) and upper ocean temperature (40–400 m; grey lines) in the sudden doubled CO_2 ensemble runs (CO2E) and transient runs (CO2T). The 21-year running means are also plotted as thick solid lines. Thick dashed lines represent the linear trends of SST and upper ocean temperature in CO2T. The temperature is averaged for the equatorial Pacific (130°E–80°W, 10°S–10°N). The mean temperature from the parallel control runs (CTRL) is removed.

response of the subsurface shows a much smaller linear trend of 0.002°C yr^{-1} than the surface (dashed grey line, Fig. 1). The equilibrium response is around 0.8°C (solid grey line), which is comparable to the surface. For the response pattern, first of all, the warming is monotonously increasing toward the high latitudes, and there is no EER (Fig. 2b). This occurs because tropical warming is impaired by the cold-water upwelling, whereas the subtropical warming is strengthened by the Ekman downwelling along the ventilation thermocline. Second, the tropical warming shows an opposite pattern to the La Niña-like warming in the surface. The reduced subsurface warming in the western tropical Pacific is resulted from the flatting of the equatorial thermocline, where the latter is caused by the weakened equatorial surface trade winds (Vecchi et al., 2006; Yang and Zhang, 2008). The flatting of the equatorial thermocline can bring the cold water upward from the lower thermocline in the west. This can be also seen in the zonal-depth section of temperature change along the equator (Fig. 3). There is clearly a weaker warming along the equatorial thermocline, whereas larger warming occurs both above and below the thermocline layer. The former is associated with radiative forcing from the atmosphere as well as the dynamic adjustment of surface oceanic circulation, whereas the latter is related to the warm water subduction along the ventilation thermocline from the subtropics as well as the lateral mixing. We will return to the mechanism in the later section.

The shallow meridional overturning circulations in the Pacific, or the subtropical cells (STCs), are weakened in response to the global warming (Fig. 4). During the transient stage with weaker CO_2 forcing, the STCs are weakened by 3% in the mass transport by the end of year 30 (Fig. 4b). The STCs are finally weakened by about 10% during the equilibrium stage under the doubled CO_2 (Fig. 4c). The slowdown of the STC results from the weakened Hadley cells (Figure not shown) because the strength of the former depends critically on the wind forcing in the subtropics (McCreary and Lu, 1994). Although its change in observations and coupled model simulations remains controversial, the Hadley cell is weakened in our global warming

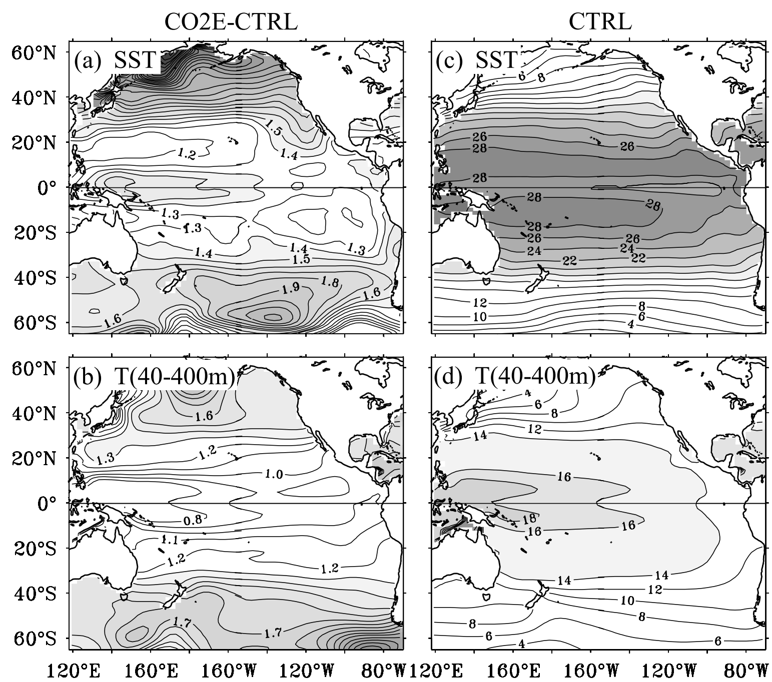

Fig. 2. (Left-hand panels) The temperature change (°C) averaged in the last 50 yr of CO2E, and (right-hand panels) the mean state of CTRL. (a) and (c) for the SST; (b) and (d) for the upper ocean (40–400 m averaged).

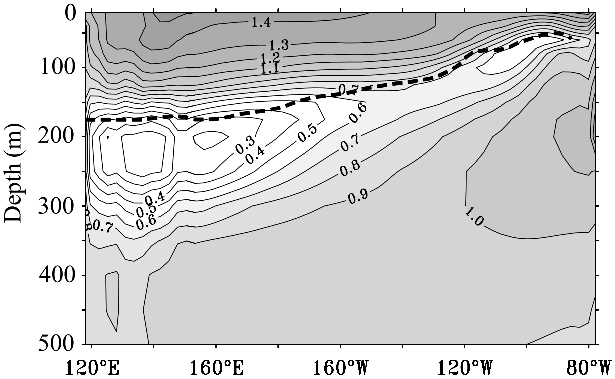

Fig. 3. Mean temperature change (°C) in the upper equatorial Pacific averaged in a 10°N–10°S strip for CO2E. The thick dashed line marks the location of the mean thermocline depth Z_{tc}. The temperature change is obtained by subtracting the mean state of CTRL from the CO2E and averaged over the last 50 yr. The Z_{tc} is defined as the location of the maximum vertical temperature gradient.

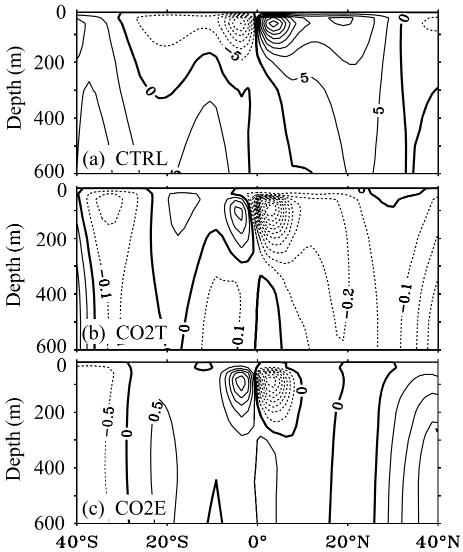

Fig. 4. Meridional overturning streamfunction of the Pacific for (a) the CTRL and changes in (b) CO2T and (c) CO2E. Contour intervals are 5 Sv in (a), 0.1 in (b) and 0.5 in (c) (1 Sv = 10^6 m^3 s^{-1}).

experiments. This change in STCs is consistent with the recent observational study (McPhaden and Zhang, 2002). It shows the slow evolution of the ocean circulation from the transient stage (Fig. 4b) to equilibrium stage (Fig. 4c).

In our experiments, the equatorial upwelling is reduced (Figs. 5a and b). It is noticed that the mean equatorial upwelling in the control run (Fig. 5a) and the anomalous equatorial downwelling in the global warming experiments (Fig. 5b) are confined within 5° of the equator. Between 5° and 10°N/S there are mean downwelling and anomalous upwelling, respectively. This is consistent with the change in the local equatorial cell. The local equatorial cell can be viewed as the equatorial component of the STC (Fig. 4). It recirculates within 10° of the equator and spirals in longitude-depth planes (Liu et al. 1994). It forms because the equatorial subsurface cold water upwells and diverges poleward by means of Ekman drift under easterlies and then descends near 10° due to Ekman convergence. In global warming, the most remarkable change in the STC occurs in the local equatorial cell (Figs. 4b and c). This is particularly clear in the western tropical Pacific (Fig. 5b). It also implies that the effect of the local oceanic dynamic feedback on the temperature should be crucial in the tropics. The stratification of the equatorial temperature is also changed significantly (Figs. 5c and d). It is enhanced in the western equatorial Pacific and weakened in the east for the upper ocean averaged between 40 and 400 m (Fig. 5d).

These mean state changes in the ocean just exhibit the consequences of the global warming. They provide no explicit answers on how and why the ocean climate should response in this way. The physical mechanisms of these changes remain unclear. In later sections, we will focus on understanding the warming patterns in the tropical Pacific. Previous studies have noticed that the warming patterns and the associated mechanisms may be different for the different stages of global warming (e.g. Yang and Zhang, 2008). In the transient stage of global warming, the atmosphere adjusts quickly to the external forcing so that it is always in a dynamic equilibrium state. The ocean, however, experiences changes that mainly concentrate near the surface because of its large thermal inertia. In the equilibrium stage with fixed external forcings, the atmosphere is almost constant whereas the ocean is still evolving, which would persist for a long time due to the slow downward penetration of the surface changes. The ocean may play very different roles in these two stages and affect differently on the coupled climate variabilities. This is important for the equatorial ocean, where the thermocline dynamics are vital to the surface climate. Here we will investigate the fundamental mechanisms of the temperature changes in different regions of the Pacific, as well as in different stages of the global warming.

4. Mechanisms

4.1. The surface

The equatorial Pacific is divided into two regions based on the different dynamics: the western equatorial Pacific (WEP: 130°E–160°W, 10°S–10°N) and the eastern equatorial Pacific (EEP: 140°W–80°W, 10°S–10°N); see Fig. 5d. The WEP includes most regions of warm pool where the surface heat flux is the dominant forcing. The EEP includes the equatorial cold tongue where the strong oceanic upwelling plays a crucial role. The mean surface climatology in the WEP is maintained by the balance between the net heat flux forcing and the meridional temperature advection ($-vT_y < 0$; Fig. 6a), whereas that in the EEP is maintained by the balance between heat flux forcing and mainly the cold-water upwelling from the subsurface ($-wT_z < 0$; Fig. 6b).

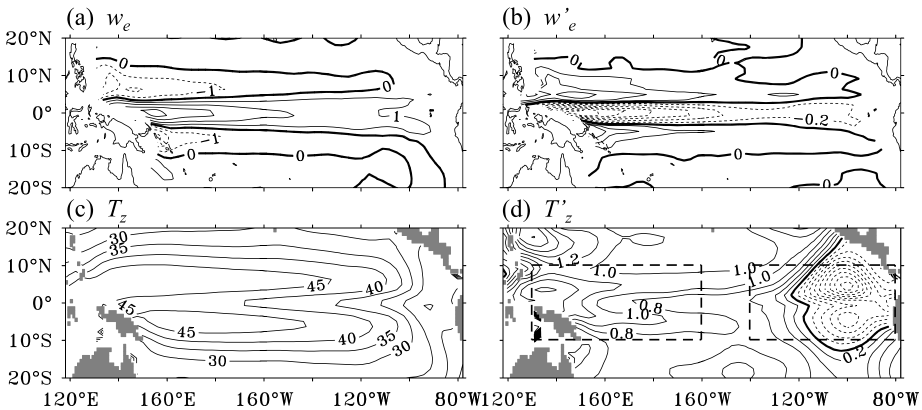

Fig. 5. (a) The subsurface mean upwelling in the CTRL and (b) its change in the CO2E averaged between 40–400 m in the tropical Pacific (unit: 10^{-6} m s^{-1}). (c) The subsurface mean vertical temperature gradient in the CTRL and (d) its change in the CO2E (unit: 10^{-3}°C m^{-1}). Contour intervals are 1 for (a), 0.2 for (b), 5 for (c) and 0.2 for (d). The dashed boxes in (d) represent the western equatorial Pacific (WEP) and eastern equatorial Pacific (EEP), respectively.

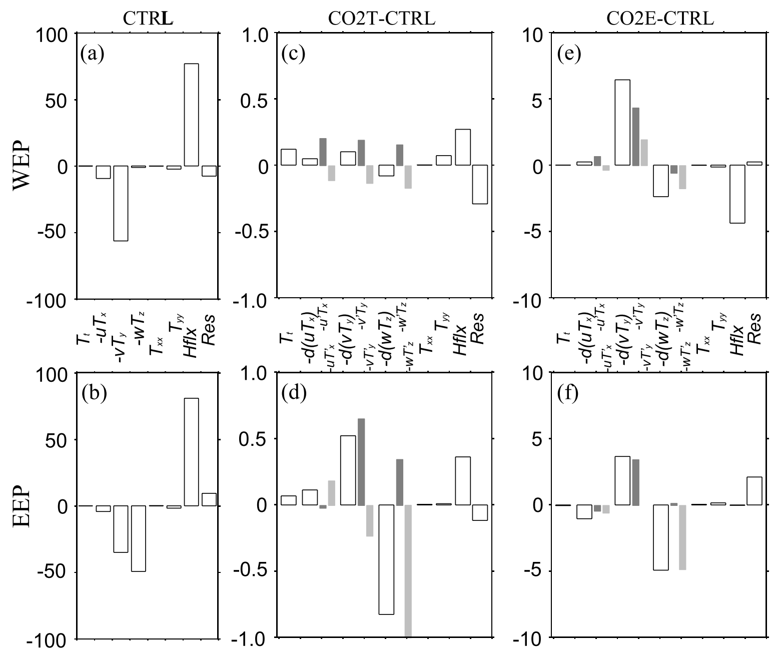

Fig. 6. The term balance of surface box of the WEP (upper panels) and EEP (lower panels). The total term balance for CTRL is plotted in (a) for WEP and (b) for EEP, averaged in year 150–200. The anomalous terms in (c) and (d) are for the CO2T and (e) and (f) for the CO2E. (unit:°C decade^{-1}). The physical meaning of each bar is marked between the upper and lower panels.

In general, the dominant mechanism to warm the surface in the transient stage of global warming is the surface heat flux, whereas that to maintain the equatorial warming in the equilibrium stage becomes the meridional advection. This is somehow out of our expectation because most of previous studies have emphasized a dominant factor of the surface heat flux or equatorial upwelling (e.g. Liu et al., 2005). Here in the equilibrium stage, the heat flux forcing becomes a damping factor instead, particularly for the WEP. For the EEP, the cold-water upwelling is the only important damping factor.

4.1.1. Transient stage. Similar to the mean climatology, under the transient CO_2 forcing the surface heat flux is the main factor to warm the equatorial SST (Fig. 6c). There are differences between the WEP and EEP. For the WEP (blank bars, Fig. 6c), the anomalous advections also contribute to the temperature increase. The vertical mixing plays an important role to restrain the surface warming. For the EEP (blank bars, Fig. 6d), the anomalous heat flux warming plays a role secondary to the strong anomalous meridional advection warming. However, the total warming effect from the three-dimensional anomalous advection is weaker than the heat flux because of the strong cold-water upwelling. The cooling effect of upwelling in the EEP is much stronger than that in the WEP, which minimizes the positive contribution of dynamic adjustment of the ocean circulation.

The net anomalous heat flux warming at the surface during the transient stage results from two processes. First, the increased atmospheric CO_2 traps infrared energy flux and eventually leads to a net anomalous long-wave heating (Liu et al., 2005). Second, because of the rapid adjustment of atmospheric circulation, the weakened surface trade wind, which is the result of slowdown of the Walker Circulation in response to the global warming (Vecchi and Soden, 2007), reduces the surface evaporation and thus the oceanic latent heat loss to the atmosphere (Yang and Liu, 2005). These two processes boost the role of the heat flux.

It is interesting to note that the oceanic dynamics start to contribute to the warming of the temperature (Figs. 6c and d), although they are still secondary to the anomalous heat flux. This is opposite to their stabilizing roles in the mean state (Figs. 6a and b). The dynamic warming of the SST arises from the changes in the ocean surface current and the equatorial upwelling as well. These changes are closely related to the change in the Pacific STCs (Figs. 4a and b). It is seen that in Figs. 6c and d, the dominant contribution to the anomalous meridional advection is due to the change in the current ($-v'T_y > 0$, black bar) instead of the change in the meridional temperature gradient ($-vT_y'$, grey bar). Usually the former is referred to as the perturbation advection mechanism (Kleeman et al., 1999; Nonaka et al., 2002; Yang and Liu, 2005), whereas the latter is referred as the mean advection mechanism (e.g. Gu and Philander, 1997). For the sea surface, the perturbation advection can also be called anomalous Ekman advection. In our experiments, both the atmospheric meridional Hadley cell and the equatorial zonal Walker circulation are weakened (figure not shown; Yang and Liu, 2005; Yang and Zhang, 2008). As a result, the strength of STC is also reduced (Fig. 4b). This causality is consistent with previous studies (McCreary and Lu, 1994; McPhaden and Zhang, 2002; Vecchi et al., 2006; Vecchi and Soden, 2007). For both the WEP and EEP, the mean meridional and vertical temperature gradient are positive (T_y, $T_z > 0$; Figs. 2c and 3); so, a weakened Ekman flow ($v' < 0$) and upwelling ($w' < 0$) result in positive perturbation advections ($-v'T_y > 0$, $-w'T_z > 0$, black bars in Figs. 6c and d) and, consequently, a rise in the local temperature. However, these positive contributions are largely cancelled by the mean advection of the anomalous temperature gradient (i.e. the mean advection mechanism; $-vT_y' < 0$, $-wT_z' < 0$, grey bars in Figs. 6c and d), particularly in the EEP where the mean upwelling ($w > 0$) is strong and the vertical temperature gradient near surface is enhanced in the transient stage ($T_z' > 0$; Fig. 1). This negative mean vertical advection impairs the warming effect of the total advections (blank bars), which in turn boosts the role of anomalous heat flux in the surface warming.

4.1.2. Equilibrium stage. Different from that in the transient stage, the surface warming in the WEP in the equilibrium stage (Fig. 6e) is maintained by the positive contribution from the anomalous meridional temperature advection ($-(vT_y)' > 0$, blank bar). The anomalous heat flux forcing has become the biggest damping factor that restrains the increase of the surface temperature instead. The anomalous vertical advection (blank bar) is the secondary damping factor. For the EEP (Fig. 6f), the surface temperature warming is also due to the positive contribution from the meridional advection, whereas the biggest damping factor here comes from the anomalous vertical advection ($-(wT_z)' < 0$, blank bar), instead of heat flux, contribution of which can be neglected in this region. It is interesting to see that the ocean dynamics, instead of the thermodynamics, plays the most important role in maintaining the warming patterns in the final stage of global warming. This is fundamentally different from the mechanisms shown in the transient stage.

As discussed in the transient stage, the dynamic control of the SST warming arises from the change in the STCs. In the equilibrium stage, the upper oceanic circulations have fully responded to the shifted surface winds. The final STC is about 10% weaker than that in control (Figs. 4a and c). This significantly enhances the positive contribution of the perturbation advection mechanism ($-v'T_y > 0$) for both the WEP and EEP (black bars, Figs. 6e and f). Physically, this can be understood as the temperature changing because less heat is transported poleward in response to the weakened Ekman flow. It is noticed that the mean advection mechanism also have considerable positive contribute to the WEP warming (Fig. 6e). Because of the EER (Fig. 7b), the anomalous meridional temperature gradient is negative ($T_y' < 0$). The poleward mean Ekman flow prevents the less warmed water poleward of 10° from cooling the WEP ($-vT_y' > 0$, grey bar, Fig. 7b). The EER is insignificant in the EEP so that the T_y' is small (Fig. 7b) and the mean meridional advection contribution is trivial here ($-vT_y' \sim 0$, Fig. 6f).

Why does the anomalous surface heat flux finally play a damping role in the SST warming in the equilibrium stage? What causes the transition of the heat flux from a positive contribution to a negative contribution? This involves negative feedback

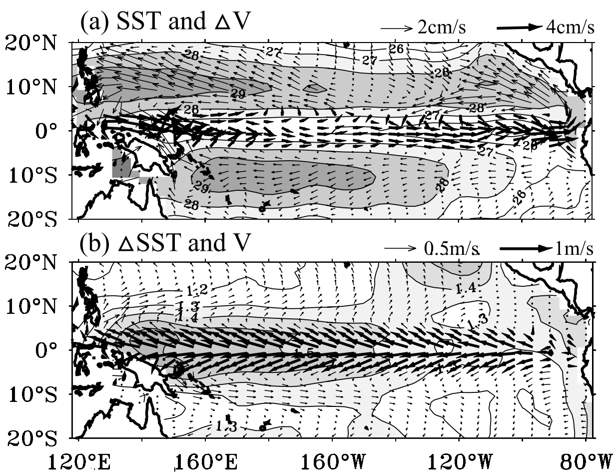

Fig. 7. (a) The mean SST (°C, shaded contours) in the CTRL and the surface current change (cm s^{-1}, vectors) in the CO2E with the mean current in the CTRL removed. (b) The mean surface current (m s^{-1}, vectors) in the CTRL and the SST change (°C, shaded contours) in the CO2E, with the mean SST in the CTRL removed.

in the air–sea interface. The change in latent heat flux is determined by two processes: change in wind speed and change in the difference of saturated and unsaturated specific humidity in the surface air. The latter change is closely related to the SST. In the transient stage, the anomalous latent heat flux is a driving mechanism to the SST warming because the weakened trade winds cause less latent heat flux lost to the atmosphere by reducing surface evaporation. However, during the equilibrium stage, the warmed ocean, in turn, increases the surface evaporation that eventually overcomes the positive effect of weakened trade winds. The latent heat flux loss is then increased and acts as a damping factor to the SST, reversing the role of the net surface heat flux in the SST warming. This transition of the role of the surface heat flux has been detailed in Yang and Liu (2005). From the point of view of balanced heat budget for the ocean (Boccaletti et al., 2004), the less poleward transport of heat requires that the ocean must lose more heat to the atmosphere to maintain a stable state. Therefore, the thermodynamic processes at the ocean–atmosphere interface are only partly responsible to the SST warming. The ocean dynamics must step in and play a crucial role in maintaining the SST warming.

The temperature change in the EEP is more dynamically controlled because the anomalous heat flux can be well neglected here (Fig. 6f). The vertical advection term ($-(wT_z)' < 0$) is the only important damping factor in the equilibrium stage. Different from the dominant role of the perturbation component in the meridional advection term, the mean vertical advection of the anomalous vertical temperature gradient ($-wT_z'$, grey bar, Fig. 6f), which is also referred to as the remote mode in some studies (Fedorov and Philander, 2001), dominates over the perturbation vertical advection ($-w'T_z$, black bar, Fig. 7f). In our experiments, the equatorial upwelling is reduced ($w' < 0$) in response to the weakened trade winds, which will result in a local warming of the SST (Fig. 6f) under the background of stable stratification ($T_z > 0$, so $-w'T_z > 0$). However, on the other hand, the vertical temperature gradient is enhanced ($T_z' > 0$, Fig. 3) because the subsurface is always less warmed than the surface, the mean upwelling ($w > 0$) tends to retard the surface warming ($-wT_z' < 0$).

Let's summarize here. First of all, the fundamental mechanism for the SST warming is different in different stages of global warming. In the transient stage, the surface heat flux forcing (or the thermodynamics) is the dominant driving factor, whereas in the stable stage, the meridional advection (or the dynamics) is the dominant driving factor. Second, the mechanism is different in the zonal direction. For the WEP, the warming is maintained by the balance between the perturbation meridional advection and the anomalous surface heat flux, whereas for the EEP the perturbation meridional advection is balanced by the mean upwelling of cold subsurface water.

4.2. *The subsurface*

For the mean subsurface climatology (Figs. 8a and b), the equatorward heat feeding from the extratropics (for the north of equator, $v < 0$, $T_y > 0$, thus $-vT_y > 0$) is the most important driving factor that is opposite to the surface. The dominant damping factor is the vertical temperature mixing for the WEP (Fig. 8a) and the vertical advection for the EEP (Fig. 8b). It is clearly seen again that the fundamental mechanism to maintain the mean temperature are different in the zonal direction.

Different from the surface warming, the mechanisms for the subsurface temperature change are consistent in different stages of global warming. Generally, the advection terms act as cooling

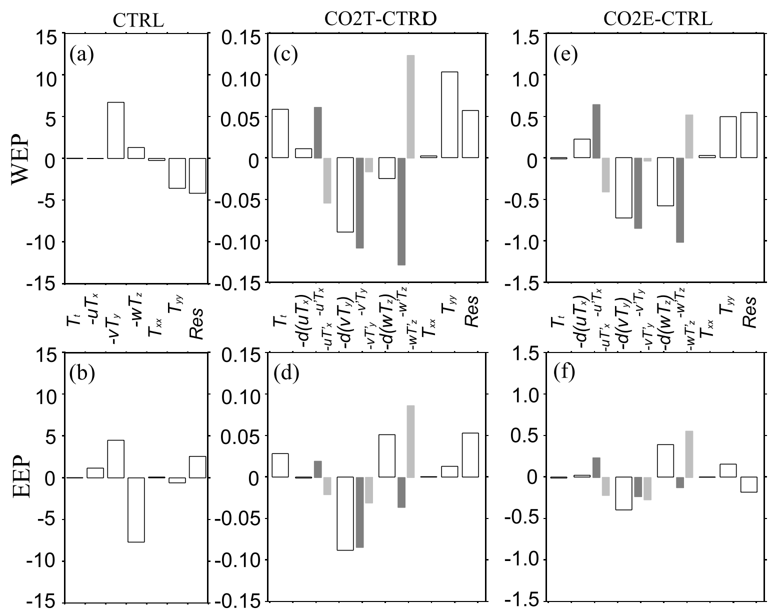

Fig. 8. Same as Figure 6, but for the subsurface box averaged between 40 and 400 m. (unit: °C decade^{-1}).

factors to the subsurface warming, which, on the contrary, boost the positive role of oceanic mixing in the subsurface warming. This is also quite different from the surface. For the WEP (Figs. 8c and e), the dominant cooling factors come from the perturbation meridional and vertical advections ($-v'T_y < 0$, $-w'T_z < 0$, black bars, Figs. 8c and e). Because of the slowdown of the STCs (Figs. 4b and c), the heat charge from the extratropics to the tropical subsurface is reduced. Although there is a strong warming effect from the mean vertical advection ($-wT'_z > 0$, grey bar), the changes in the ocean vertical and horizontal circulations ($-v'T_y$, $-w'T_z$) are more important than the mean warm-water advection. From the transient (Fig. 8c) to the equilibrium stage (Fig. 8e), the contribution from the circulation change is becoming bigger. This is consistent with the changes in STCs and manifests the slow evolution of the ocean circulation. Different from the WEP, in the EEP the vertical advection is a warming factor to the subsurface (Fig. 8d and f). This occurs because the dominant role of the anomalous warm water upwelling ($wT'_z > 0$, grey bar). The thermocline in the EEP is shallow and the ocean is warmed from both the upper and lower side of thermocline (Fig. 3). The vertical average of the anomalous temperature gradient between 40–400 m results in a negative T'_z (Fig. 5d). On the other hand, the anomalous upwelling in the EEP is much smaller than in the WEP (Fig. 5b); therefore, relatively the perturbation upwelling cooling effect ($-w'T_z < 0$, black bar, Figs. 8d and f) is much weaker than the warming effect of the mean upwelling of anomalous warm water.

It is worth noting the positive contributions of the meridional and vertical mixings to the subsurface warming. The pattern of the temperature warming in the tropics (Figs. 2 and 3) indicates that the ocean stability in the upper ocean, particularly in the WEP (Figs. 3 and 5d), is increased throughout the entire global warming period. This suppresses the vertical diffusivity through a Richardson number-dependent parametrization of the vertical diffusivity in the ocean model (Pacanowski and Philander 1981) and, in turn, the cold diffusive flux and entrainment, resulting in an anomalous warm vertical diffusion (Figs. 8c–e; Liu et al., 2005). In the EEP, the warming of subsurface in the equilibrium stage in turn reduces the ocean stability (Fig. 5d), resulting in a strong, cold vertical diffusion and entrainment and, thus, an anomalous cold impact (Fig. 8f). The warming effect from the meridional diffusion term is also very strong, particularly for the WEP (Fig. 8c), consistent with the change in meridional temperature distribution ($T'_y > 0$; Fig. 9b). Since the changes in the large-scale ocean circulation tend to cool the subsurface, the subsurface warming has to rely on the small- or microscale turbulent mixing. This is again quite different from the conventional point of view that emphasizes the role of the meridional advection or subduction along the isopycnal level in the tropical–extratropical climate interaction (Gu and Philander, 1997). The meridional turbulent diffusion, at least based on our study here, is very important to the tropical thermocline climate change on the decadal and longer timescale. The extraradiative

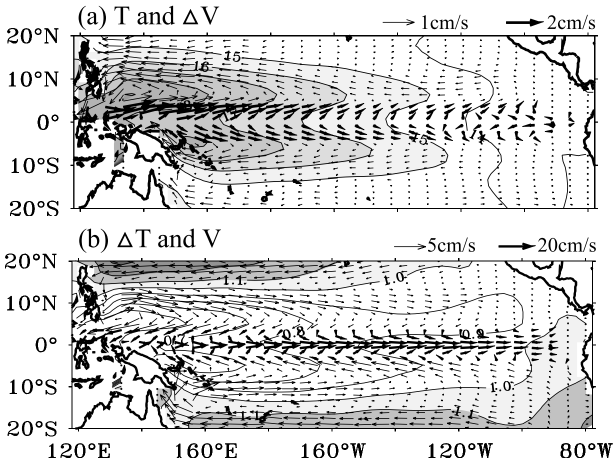

Fig. 9. Same as Figure 7, but for the subsurface ocean averaged between 40 and 400 m.

atmospheric heating caused by unbalanced GHGs' effect has to be dissipated into the deeper ocean mainly through the turbulent diffusion processes.

5. Conclusions and discussions

This work investigates the physical processes that are responsible for the ocean temperature changes in global warming. Because of the slow evolution of the ocean, we found that the dominant mechanisms for the ocean temperature changes are different in different stages of global warming. By analysing temperature budget in transient and equilibrium CO_2 experiments, we find that the dominant mechanism to warm the tropical Pacific surface in the transient stage is the surface heat flux, whereas that to maintain the equatorial warming in the equilibrium stage is the meridional advection. This differs from previous studies that emphasize the dominant factor of the surface heat flux or equatorial upwelling. Different from the surface warming, the mechanisms for the subsurface temperature change are consistent in the entire period of global warming. It is found, surprisingly, that the subsurface warming has to rely on the small- or microscale turbulent mixing instead of the warm-water subduction from the extratropics as conventional belief. The meridional turbulent diffusion appears to be critical to the tropical thermocline climate change on the decadal and longer timescale.

The transition of the role of surface heat flux in the tropical SST warming actually results from the dynamic adjustment of the ocean circulation to the global warming. In our coupled model, both the atmospheric Hadley cell and Walker circulation are slackened in response to the global warming, so are the ocean circulations, such as the STCs in the Pacific. This results in the reduced polewards heat transport. From the point of view of balanced heat budget for the ocean (Boccaletti et al., 2004), the less poleward heat transport requires that the ocean must lose more heat to the atmosphere to maintain a stable state.

Therefore, we would like to emphasize the deterministic role of the dynamic adjustment of the ocean circulation in the long-term climate change.

It is worth noting that the physical processes are quantitatively investigated in this paper by explicitly calculating their contributions. The critical role of the turbulent mixing in the tropical subsurface temperature change is pinpointed by these quantitative analyses. This contradicts the previously plausible point of view that emphasizes the critical role of the extratropical water subduction in maintaining the thermal structure of the tropical thermocline. Obviously, our arguments may be questionable and need to be examined strictly in other models and by other approaches.

The response patterns and associated mechanisms examined in this paper are appropriate for our model and could be model-dependent. Our results can be viewed as one manifestation of a broad spectrum of possibilities, and thus, the study here should enhance our understanding of the real world and provide clues to further identify and understand the possible future climate changes, although it may bring up more disputes than consent. This work only focuses on the tropical Pacific. Detailed analyses on other parts of Pacific and other oceans are right on the way.

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7. Appendix A

Let's start from the temperature equation

$$\frac{\partial T}{\partial t} = -uT_x - vT_y - wT_z + A_h T_{xx} + A_h T_{yy} + Q_{\mathrm{F}} + R_{\mathrm{E}}. \tag{A1}$$

Here, $\partial T/\partial t$ is the local temperature tendency, $-uT_x$, $-vT_y$ and $-wT_z$ are the zonal, meridional and vertical temperature advection, respectively, $A_h T_{xx}$ and $A_h T_{yy}$ are the horizontal diffusion terms with the constant diffusion coefficients A_h (4000 $\mathrm{m^2\ s^{-1}}$), Q_{F} is the surface net heat flux forcing (it is zero for the subsurface). The residual term R_{E} includes the vertical diffusion term and the vertical convection, which are obtained by subtracting the other terms from $\partial T/\partial t$, since they cannot be explicitly calculated.

For the equilibrium stage of global warming with a fixed doubling CO_2, the changes of terms in (A1) are derived as the differences of the terms in the CO2E experiments and their control runs. These anomalous terms are,

$$\partial \mathrm{T}'/\partial \mathrm{t} = -(uT_x)' - \left(vT_y\right)' - (wT_z)' + \mathrm{A}_h T'_{xx} + \mathrm{A}_h T'_{yy} + (Q_{\mathrm{F}})' + (R_{\mathrm{E}})', \tag{A2}$$

where, $\partial T'/\partial t = (\partial T/\partial t)_{\mathrm{CO2E}} - (\partial T/\partial t)_{\mathrm{CTRL}}$ and all other terms are similarly obtained. In the paper, the time average of the last 50-year simulation is analysed.

For the transient stage of global warming, the contribution to temperature change by a specific forcing F in (A1) is first calculated (Liu et al., 2005) for both the CO2T and control simulations separately,

$$T_F(x, y, t) = \int_o^t F(x, y, t)\,\mathrm{d}t. \tag{A3}$$

Then, the linear trend of this contribution T_F is calculated. Finally, the difference between the linear trends of the two simulations is taken as the anomalous heat budget responsible for the transient CO_2-induced temperature change, similar as (A2).

The temperature advection terms can be further decomposed as

$$\begin{aligned} -(uT_x)' &= -u'T_x - uT'_x - u'T'_x, \\ -\left(vT_y\right) &= -v'T_y - vT'_y - v'T'_y, \\ -(wT_z)' &= -w'T_z - wT'_z - w'T'_z. \end{aligned} \tag{A4}$$

Thus, the changes in the temperature advection result from the changes of mean current ($-u'T_x$, $-v'T_y$, $-w'T_z$) and mean temperature gradient ($-uT'_x$, $-vT'_y$, $-wT'_z$) as well as their non-linear interactions ($-u'T'_x$, $-v'T'_y$, $-w'T'_z$). The temperature gradient is positive poleward and upward. Usually the non-linear terms are so small that they can be safely discarded. We will not discuss the non-linear terms in the paper.

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