Mechanisms of Fast Walker Circulation Responses to CO2 Forcing

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Abstract

The Walker circulation (WC) responds to CO\textsubscript{2} forcing at both short and long timescales. The fast response accounts for a substantial portion of the total responses, but its mechanisms, particularly those pertaining to air-sea interactions remain unclear. We found the initial land-sea warming contrast drives the anomalous easterlies (westerlies) in the Pacific warm pool (cold tongue) region. When the anomalous easterlies dominate, a Bjerknes feedback is triggered, leading to fast Equatorial Pacific cooling and WC strengthening. Otherwise, the Central and Eastern Pacific warms via the wind-SST-evaporation feedback and a weakened upwelling, leading to WC weakening. While the WC weakens and the equatorial Pacific warms eventually, the transition into the slow response differs among climate models. If initial cooling happens, the Equatorial warming emerges as the Bjerknes feedback wanes and the subsurface warm water accumulated in the warm pool migrates eastward via downwelling oceanic Kelvin waves.
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Key Points:

• Climate models disagree on fast Pacific Walker circulation responses to \textit{CO}_2 forcing.
• The disagreement results from a tug of war between anomalous easterlies in the Pacific warm pool and anomalous westerlies in the cold tongue.
• When anomalous easterlies dominate, the Walker circulation strengthens initially, and then weakens as the climate warms.
Abstract

The Walker circulation (WC) responds to CO$_2$ forcing at both short and long timescales. The fast response accounts for a substantial portion of the total responses, but its mechanisms, particularly those pertaining to air-sea interactions remain unclear. We found the initial land-sea warming contrast drives the anomalous easterlies (westerlies) in the Pacific warm pool (cold tongue) region. When the anomalous easterlies dominate, a Bjerknes feedback is triggered, leading to fast Equatorial Pacific cooling and WC strengthening. Otherwise, the Central and Eastern Pacific warms via the wind-SST-evaporation feedback and a weakened upwelling, leading to WC weakening. While the WC weakens and the equatorial Pacific warms eventually, the transition into the slow response differs among climate models. If initial cooling happens, the Equatorial warming emerges as the Bjerknes feedback wanes and the subsurface warm water accumulated in the warm pool migrates eastward via downwelling oceanic Kelvin waves.

Plain Language Summary

The Walker circulation (WC), which describes the large-scale tropical air flow in zonal and vertical directions, is a product of the interaction between the atmosphere and ocean. A strengthening of the WC is associated with the Bjerknes feedback: as the westward trade winds keep warm water in the western Pacific and draw up deep, cold water in the eastern Pacific, the zonal sea surface temperature (SST) gradient is enhanced. The zonal SST gradient reinforces trade winds as they blow from east to west. There is no clear census whether the WC will strengthen or weaken under global warming. We use multiple climate models to study the response of WC to CO$_2$ forcing at different timescales. Immediately following the forcing, we found two types of responses among climate models and the difference between the models is set by how surface winds change initially. An intensification of trade winds in the western Pacific strengthens the WC and cool the Equatorial Pacific. On the other hand, a weakening of trade winds in the eastern Pacific weakens the WC and warm the Equatorial Pacific. The relative strength of these two processes determines how the WC changes initially and its transition into the long-term response.

1 Introduction

The response of the Walker circulation (WC) to greenhouse gas forcing has a significant effect on tropical climate and the hydrological system. Substantial research has been done on WC changes, yet it is not entirely clear whether the WC will strengthen (L’Heureux et al., 2013; Kohyama et al., 2017), weaken (Held & Soden, 2006; Vecchi et al., 2006; Heede et al., 2020) or remain unchanged under continued global warming. Most climate models project a weakening of the WC as the mean sea surface temperature (SST) increases (Vecchi et al., 2006; Held & Soden, 2006; Seager et al., 2010; He & Soden, 2015). In observations, the strength of the WC has intensified over the last half of the twentieth century, but such a trend is substantially influenced by internal variability (Chung et al., 2019; Meng et al., 2012).

Previous studies have suggested that both the direct CO$_2$ forcing and subsequent surface temperature changes, including the land-sea warming contrast, the global mean warming and the pattern of sea surface temperature (SST) changes, play a role in the WC response (Held & Soden, 2006; He & Soden, 2015). In the real world, where the WC responds to concurrent changes in CO$_2$ concentration and SST, it’s difficult to determine the relative roles of the various forcing factors. Abrupt forcing experiments have been commonly used to dissect the various mechanisms based on their distinct timescales (e.g., Andrews et al. (2009); Chadwick et al. (2019)). In such experiments, the response of the WC exhibits a fast and a slow component. The slow component consists of a weakening of the WC driven primarily by the gradual warming of the global mean SST (e.g., Held...
and Soden (2006); He and Soden (2015)). The fast component, however, is more complicated. On the one hand, the direct CO$_2$ forcing weakens the WC through the stabilization of the troposphere (Bony et al., 2013; He & Soden, 2015; Thorpe & Andrews, 2014); on the other hand, the land-sea warming contrast tends to strengthen the WC through the increase of convection over tropical land (Kamae & Watanabe, 2013; He & Soden, 2015). The pattern of SST changes may also play a role, as changes in convection and SST are spatially correlated at both fast and slow timescales (Chadwick et al., 2014; He & Soden, 2015). However, details of the interaction between the WC and SST during the fast response period remain unclear.

Given the fact that the fast response accounts for a substantial portion of the total WC response (Bony et al., 2013; Chadwick et al., 2014; He & Soden, 2015), it is important to understand what drives the fast WC response, as well as the transition to the slow response. This study investigates the underlying mechanisms of the fast WC response with a focus on the role of air-sea interactions. We found both an initial strengthening and weakening of the WC in different models, each associated with a distinct transition to the slow response.

2 Methods

2.1 Fast versus slow

The total response of the WC to CO$_2$ forcing contains a fast and a slow components. The fast response involves changes and feedbacks directly triggered immediately by increasing CO$_2$. The slow response, on the other hand, scales with the gradual warming of the global mean SST warming (Bony et al., 2013; Samset et al., 2016). There are two common approaches to isolate the fast circulation response: 1) fixed SST experiment, and 2) fully coupled experiment with an abrupt CO$_2$ forcing (Ceppi et al., 2018; Chadwick et al., 2014; He & Soden, 2016; Kamae & Watanabe, 2013; Merlis, 2015; Richardson et al., 2016). By increasing CO$_2$ while fixing SST in atmosphere-only models, the first approach isolates the part of fast response caused by the direct CO$_2$ forcing and land-sea warming contrast, but it does not allow for interactions between changes in circulation and SST. The second approach separates the fast and slow response based on their time of emergence, while the atmosphere and SST are coupled at both fast and slow timescales. In order to understand the importance of air-sea coupling during the fast response, we use fully coupled abrupt 4×CO$_2$ experiments in this study. Particularly, we define the fast response based on the evolution of changes during year 1 to year 2 after the CO$_2$ quadrupling, the total response as the average change of year 91 to 120, and the slow response as the difference between the total and the fast response. We calculate the percentage of the fast to the total response as following:

$$\text{Fast response} \% = \frac{|\text{Fast response}|}{|\text{Fast response}| + |\text{Slow response}|} \times \%$$ (1)

2.2 Definition of Changes in Walker Circulation Strength

The WC strength is calculated as the difference in 500 hPa pressure velocity (Ω in hpa) between the Indian-West Pacific (IWP, regions between 50°E−150°E, 10°S−10°N) and the Central-East Pacific (CEP, regions between 210°E−270°E, 10°S−10°N). We concern the anomalies normalized by the corresponding long-term pre-industrial control climatology (Eq. 2). The changes in the WC strength are expressed as:

$$\Delta\text{WC strength} = \frac{\Delta(\Omega_{500}(\text{IWP}) - \Omega_{500}(\text{CEP}))}{\text{control}(\Omega_{500}(\text{IWP}) - \Omega_{500}(\text{CEP}))}$$ (2)
2.3 Model Simulations

We analyze the abrupt $4 \times CO_2$ experiment from phase 5 of the Coupled Model Intercomparison Project (CMIP5) (Taylor et al., 2012) and a large ensemble abrupt $4 \times CO_2$ experiment of Community Earth System Model version 1 (CESM1) (M. A. A. Rüggenstein et al., 2016). The CESM1 $4 \times CO_2$ large ensemble (LE) consists of 120 ensemble members for the first two years and three ensembles for years 91 to 120. All CMIP5 models (27 in total) that provide all necessary variables are used.

As shown in Fig. 1a and Fig. S1, the fast response is diverse among CMIP5 models. Such diversity is partially associated with internal variability acting within the first two years. To minimize the effect of internal variability, models are divided into a cold group (CG) and a warm group (WG) based on their initial SST changes in the Niño 3.4 region (Fig. 1a and Fig. S1): models with values lower than the multi-model ensemble mean (CMIP5 MEM) are grouped into the CG and the rest into the WG. The CG models on average simulate an initial cooling or less warming in the Central Pacific and strong anomalous easterlies in the Pacific warm pool, whereas the WG models on average simulate an enhanced initial warming in the Central Pacific and strong westerlies in the Pacific cold tongue (Fig. S1).

2.4 Robustness

We determine the robustness of changes by quantifying the range of internal variability using the Monte Carlo method. The long-term pre-industrial control experiments of each model are used. Here we describe the quantification of the range of internal variability associated with 6-month changes averaged over N ensembles as an example (N corresponds to the number of CMIP5 models or the number of CESM1 ensemble members). We first calculate the difference between two non-overlapping 6-month periods randomly sampled from the control run. This is done N times and the average difference is taken as one sample of internal variability. This process is repeated 5000 times to produce a probability distribution of the internal variability for 6-month changes averaged over N ensembles. Ninety-five percent confidence level is shown in Fig. 1 based on the Student’s t-distribution.

3 Results

3.1 Fast response

In the first two years after abrupt $CO_2$ quadrupling, the CESM1 LE and the CMIP5 CG models show an average cooling in the central Equatorial Pacific and a warming in the Pacific cold tongue region (Figs. 2a and 2c). The WC strengthens and anomalous easterlies form in the Pacific warm pool region (Figs. 1c and 1e). The cooling of the central Equatorial Pacific and the strengthening of the WC are connected via the Bjerknes feedback: Within the first two months after quadrupling $CO_2$ and before a substantial SST response, a land-sea warming contrast develops and drives the anomalous easterlies in the warm pool region (Fig. S2 a). The land-sea warming contrast drives the convection over the Maritime Continent (MC), which intensifies the anomalous easterlies to the west of the MC as the Kelvin wave response to the anomalous convective heating (Fig. S3; e.g., Gill (1980)). The anomalous easterlies increase the zonal SST gradient in the Equatorial Pacific, which in turn strengthens the wind anomalies via the Bjerknes feedback. A similar Bjerknes feedback appears in the Equatorial Indian Ocean, where the anomalous westerlies and the west-east SST dipole reinforce each other, and together further intensify the anomalous convection in the MC (Figs. S3 a and d; Wang et al. (2016)). As a result of the feedbacks between changes in the circulation and SST, the WC strengthens and is shifted westward (Figs. 1c, 3a, and 3c). On the other hand, the land-sea warming contrast drives anomalous westerlies in the Pacific cold tongue (Figs. 1d, 2a, 2c, 2e)
and S2 a). The anomalous westerlies suppress the coastal Ekman upwelling, warming the Pacific cold tongue. The fast SST and circulation responses differ substantially from the total responses in the CESM1 LE and CG models (Figs. 2a - 2d), indicating a non-monotonic evolution of Indo-Pacific climate under CO2 forcing.

The CMIP5 WG models, however, simulate a virtually monotonic response. As shown in Fig. 2e, an enhanced warming develops in the eastern equatorial Pacific within the first two years, which is somewhat similar to the total response (Figs. 2e and 2f). The initial surface wind response contains westerly anomalies in the Pacific cold tongue region driven by the fast warming of the South American Continent but virtually no easterly anomalies in the Pacific warm pool. The fast warming of the eastern equatorial Pacific in the WG models is likely driven by two mechanisms: 1) via the reduced evaporation driven by the weakening of the trade winds (i.e., a wind-evaporation-SST feedback, e.g., X. Li et al. (2016)), and 2) the weakening of the Ekman upwelling in the Pacific cold tongue region (Heede et al., 2020).

What causes the disparity in the fast responses between the CG models (including CESM1 LE), and the WG models? As shown in Figs. 1d and 1e, the magnitude of the anomalous westerlies in the Pacific cold tongue regions are similar between the two model groups, but the CG models have a much stronger anomalous easterlies in the Pacific warm pool regions, namely 2-3 m/s larger than the WG models. The fast response is determined by a tug of war between the anomalous easterlies and westerlies. When the anomalous easterlies dominate, the positive Bjerknes feedback is triggered, cooling the Equatorial Pacific in the CG models. While the anomalous westerlies act to warm the Equatorial Pacific from the east, the effect is restricted to a small region near the coast of South America. In the WG models, however, the anomalous westerlies dominate, leading to a fast Equatorial warming that is similar to the total response. Our suggested mechanism is further supported by changes in the subsurface ocean temperature (Figs. 4a, 4e and 4i). The anomalous westerlies suppress the coastal Ekman upwelling, warming the upper ocean in the cold tongue region in both model groups. On the other hand, surface wind changes in the western Pacific act to deepen the thermocline in the CG models and flatten it in the WG models. We do not fully understand why different groups simulate different anomalous easterlies during the fast response, but we have the following two hypotheses: The WG models simulate much less initial convection changes over the MC compared to the CG models (compare Figs. S3 a, d and g), which could result from different model parameterizations. Another explanation is that WG models have less sensitivity of anomalous easterlies to the changes in MC convection when compared to CG models (Fig. S7). We also tested the relationship between the initial land-sea warming contrast and anomalous surface wind in the warm pool region. As shown in Fig. S8, in the first three months when the Bjerknes feedback has not formed, there is no strong correlation between initial land-sea warming contrast and anomalous easterlies among 27 CMIP5 models.

Even though the CG models behave quite differently than the WG models in the fast response period, the fast response accounts for a substantial portion of the total response for both model groups. The percentage of the fast response is roughly 45% for central Pacific SST and 30 ~ 60% for WC strength (texts on Figs. 1b and 1b). If we look at the zonal mean pattern of WC (Figs. 3g-3i), the fast response actually accounts for more than 80% over the Maritime Continent (around 100°E-120°E) for CG models, and more than 60% over the equatorial Atlantic basin (around 60°W-10°W) for all both WG and CG models.

### 3.2 Transition from the fast to the slow response

Despite the disparity in the fast response, the slow response is qualitatively similar across model groups. The total response features a weakening of the WC (Fig. S5;
Vecchi and Soden (2007); He and Soden (2015)), and an enhanced Equatorial Pacific warming (Fig. S4; DiNezio et al. (2009); Xie et al. (2010); G. Li et al. (2016); Heede et al. (2020); M. Rugenstein et al. (2020)) in both model groups. The amplitude of the total response is larger in the WG models because of the lack of Bjerknes feedback during the fast response period (Figs. 1b and 1c). The slow response includes an anomalous upward motion over the central and eastern Equatorial Pacific and an eastward shift of the WC (Fig. S5). The mean SST warming and the pattern of SST warming are two main drivers of the slow response (He & Soden, 2015). As shown in Fig. S5, the mean SST warming weakens the convection over the Indo-west Pacific. This weakening mechanism can be explained by two mechanisms. The first one is the advection of stratification change effect, which produces anomalous air sinking (uplifting) in the climatological ascending (descending) regions, in respect of the dry stability increase (Knutson & Manabe, 1995; Ma et al., 2012).

The second one is because the global mean precipitation increases faster than the global mean atmospheric moisture, which is associated with an overall weakening of the vertical motions (Held & Soden, 2006; Vecchi & Soden, 2007). The pattern of SST warming, particularly the enhanced Equatorial Pacific warming, contributes to the weakening of the circulation by shifting the WC eastward (He & Soden, 2015). For the WG models, much of the effect of the pattern of SST warming is partially achieved during the fast response (compare Figs. 2e and 3c to Figs. S4 c and S5 c). During the slow response, the zonal SST gradient continues to weaken due to the advection of extratropical warm water, further weakening the WC over time (Luo et al., 2017; Heede et al., 2020). As shown in Figs. 3g-3i, both the fast and slow responses account for a substantial portion of the total WC response, which is consistent with He and Soden (2015). The sign of the total WC response is determined by the slow response, whereas much of the inter-group discrepancy can be attributed to the fast response.

The transition from fast to slow response for CMIP5 WG models is homogeneous, as the WC continues to weaken as the mean SST increases (Figs. 3c and S5 c). The transition for the CESM1 LE and CMIP5 CG models starts with the weakening of convection in the MC (Fig. S3). This could be associated with the overall weakening of tropical convection as the global mean SST increases (He & Soden, 2015) or caused by the diminishing effect of the land-sea warming contrast as the low-level moisture static energy slowly homogenizes between land and ocean (Byrne & O’Gorman, 2013). As the convection weakens, so do the anomalous easterlies in the western Pacific warm pools (Figs. 1d, 2b and 2d). The Bjerknes feedback, which maintains the Equatorial cooling during the fast response, starts to wane. The initial warm water anomalies that are accumulated in the warm pool by the anomalous easterlies (Figs. 4a and 4e) begin to migrate eastward as downwelling Kelvin waves. Given the speed of oceanic Kelvin waves (~ 2 to 3 m/s), it should take about 4 months for the wave front to travel across the Pacific (Roundy & Kiladis, 2006). Indeed, it takes less than a year for the surface cooling to disappear, which occurs between years 4 and 5 in the CESM1 LE (Figs. 4b and 4c) and around year 2 in the CMIP5 CG models (Fig. 4f). As the cooling pattern disappears, the CESM1 LE and the CMIP5 CG models eventually develop an enhanced Equatorial warming similar to that in the WG models (Figs. 4d, 4h and 4i).

4 Summary and discussion

We find two distinct fast responses of the Walker circulation to CO2 forcing in the climate models. Whether the WC strengthens or weakens initially is determined by a tug of war between the anomalous westerlies in the eastern Equatorial Pacific and the anomalous easterlies in the western Equatorial Pacific, both driven by the initial land-sea warming contrast. In models where the anomalous easterlies overpower the anomalous westerlies, a cooling pattern appears in the central Equatorial Pacific, leading to an initial strengthening and westward shift of the WC. As the global mean SST slowly increases, the anomalous easterlies gradually disappear. The Equatorial Pacific starts to
warm as the warm water accumulated initially in the Pacific warm pool travels eastward as Kelvin waves, and the WC eventually weakens. In models where the anomalous westerlies dominate, the central and eastern Equatorial Pacific warm initially, leading to a fast weakening of the WC. In both model groups, the slow increase of the global mean SST eventually weakens the WC, which persists over time.

As shown in Figs. 1 and 3, the CG models on average project less total WC weakening than the WG models. Such a discrepancy results from the disparity in the fast WC response, as the CG models actually simulate more slow WC weakening on average (compare Figs. S5 b and S5 c). We hypothesize that the inter-model spread in the WC response may be largely attributed to the spread in the fast response. To fully test such hypothesis, large ensemble of abrupt $4 \times CO_2$ simulations from individual models are required. Although the large ensemble technique has been widely used as a means to separate internal and anthropogenic climate changes (Kay et al., 2015; Deser et al., 2020), only few modeling centers have abrupt $4 \times CO_2$ type large ensemble experiments. Here, our results suggest that a greater application of abrupt $4 \times CO_2$ large ensembles would be helpful to better understand the uncertainty in the WC changes.

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### References


L’Heureux, M. L., Lee, S., & Lyon, B. (2013). Recent multidecadal strengthening of
the walker circulation across the tropical pacific. Nature Climate Change, 3(6), 571–576. Retrieved from https://doi.org/10.1038/nclimate1840 doi: 10.1038/nclimate1840


Figure 1. Separation of CMIP5 CG (blue) and WG (red) groups based on the changes in SST over Niño 3.4 region during fast response period (a). Changes in SST over the central Pacific (190°E-240°E and 5°S-5°N) subtracted by the changes in SST over the entire tropical (30°S-30°N) mean (b), changes in the WC strength which is defined in Section 2.2 (c), changes in zonal surface wind in the cold tongue (regions between 240°E-270°E, 2°S-2°N) (d) and the warm pool (regions between 150°E-180°E, 2°S-2°N)(e) at different time periods after abrupt 4×CO₂. The orange bar in plot (a) indicates the multi-model ensemble mean and ΔSST is normalized by the tropical mean change of surface temperature. In plot (c), positive values indicate the strengthening of Walker circulation and vice versa. The semi-transparent striped areas in plots (b) to (e) indicate the 95% confidence level based on the probability distribution of models’ internal variability (Section 2.4). The texts on plots (b) to (e) are the percentages of the fast response to the total response of different model groups (Eq. 1).
Figure 2. Fast (a, c, e) and total (b, d, f) response of TS (shadings) and surface wind (vectors) for CESM1 LE (a, b), CMIP5 CG (c, d), and CMIP5 WG (e, f). $\Delta$TS is normalized by the entire tropical mean ($30^\circ$S-$30^\circ$N) change of surface temperature.
Figure 3. The fast response (a-c) and total (d-f) response of zonal mean mass stream functions with an interval of the $5 \times 10^{10}$ kg s$^{-1}$ (shadings) and vertical wind (vectors) for CESM1 LE (a, d), CMIP5 CG (b, e), and CMIP5 WG (c, f). The vertical pressure velocity is scaled by a factor of 300. The contours show the control climatology of the zonal mean mass stream functions with an interval of $10^{10}$ kg s$^{-1}$. The percentages of the fast response of the zonal mean mass stream functions for each group are presented in plots g-i.
Figure 4. Time evolution of equatorial Pacific subsurface ocean temperature (averaged over 5°S-5°N) for CESM1 LE (a, b, c, d), CMIP5 CG (e, f, g, h), and CMIP5 WG (i, j, k, l). Year 4 to 6 are plotted for CESM1 LE, while Year 2 to 4 are plotted for CMIP5. We choose these specific time intervals to show how the downwelling Kelvin oceanic waves erase the initial Pacific cooling. The ocean temperature warms monotonically after year 6 for all models (See Fig. S6 for the slow response of the subsurface ocean temperature).
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Contents of this file

Figures S1 to S8

Introduction

This supporting information includes supplementary Figure S1 to S8. Figure S1 shows the changes in surface wind and surface temperature for 27 CMIP5 models after quadrupling CO$_2$. Figure S2 demonstrates the fast TS and surface wind changes in CESM1 LE. Figure S3 shows the changes in 500hPa pressure velocity in the first 2 years, year 5 to year 15, and year 91 to year 120 after quadrupling CO$_2$. Figure S4-S6 show the slow response of TS, surface wind change, the Walker circulation and subsurface ocean temperature. Figure S7 shows the relationship between anomalous easterlies in the warm pool region and convection over MC. Figure S8 demonstrates the relationship between land-sea warming contrast and anomalous easterlies in the warm pool region.
Figure S1. Fast response of TS (shadings) and surface wind (vectors) for 27 CMIP5 models. ΔTS is normalized by the equatorial mean change of surface temperature. The name of individual model and its ΔSST in Niño 3.4 region are labeled on the top left of each plot. CMIP5 cold group (CG) models are: ACCESS1-0, CanESM2, CNRM-CM5, GFDL-ESM2M, IPSL-CM5A-MR, IPSL-CM5A-LR, MPI-ESM-P, NorESM1-ME, NorESM1-M, CCSM4, BNU-ESM, FGOALS-s2, CNRM-CM5-2 and inmcm4. CMIP5 warm group (WG) models are: ACCESS1-3, bcc-csm1-1-m, GISS-E2-H, GISS-E2-R, HadGEM2-ES, IPSL-CM5B-LR, MIROC5, MIROC-ESM, MPI-ESM-MR, MRI-CGCM3, bcc-csm1-1, CSIRO-Mk3-6-0 and MPI-ESM-LR.
Figure S2. Evolution of surface temperature and surface wind changes in the first two years after abrupt 4 × CO2. We use CESM LENS instead of CMIP5 CG results here because CESM LENS has 120 ensembles for each month, while CMIP5 CG only has 14.

Figure S3. Changes in 500 hPa pressure velocity (shadings) and surface winds (vectors) for CESM1 LE (a, b, c), CMIP5 CG (d, e, f) and CMIP5 WG (g, h, i).
Figure S4. Slow response of TS (shadings) and surface wind (vectors) for CESM1 LE, CMIP5 CG and CMIP5 WG. $\Delta TS$ is normalized by the tropical mean change of surface temperature.

Figure S5. Slow response of zonal mean mass stream functions with an interval of $5 \times 10^{10} \text{ kg s}^{-1}$ (shadings) and vertical wind (vectors) for CESM1 LE, CMIP5 CG and CMIP5 WG. The vertical pressure velocity is scaled by a factor of 300. The contours show the control climatology of the zonal mean mass stream functions with an interval of $10^{10} \text{ kg s}^{-1}$. 
**Figure S6.** Slow response of equatorial Pacific subsurface ocean temperature (averaged over 5°S — 5°N) for CESM1 LE, CMIP5 CG and CMIP5 WG.

**Figure S7.** Changes in anomalous easterlies in the warm pool (region between 2°S - 2°N and 150°E - 180°E) plotted against changes in pressure velocity at 500hpa over MC region, which is defined as the region between 10°S - 10°N and 90°E - 110°E for CESM1 LE (blue), CMIP5 WG (red) and CMIP5 CG (green). The lines are the least square fit to the data and the numbers (in the legend) are the slopes of each line.
Figure S8. Changes in anomalous easterlies of the first 3 months after abrupt CO₂ in the warm pool (region between 2° S - 2° N and 150° E - 180° E) plotted against the land-sea warming contrast. The land-sea warming contrast index is defined as the temperature difference between all lands and ocean over 15° S - 15° N. The Bjerknes feedback has not formed yet in the first three months.