Asymmetric impact of the North and South Pacific on the equator in a coupled climate model

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[1] Relative contributions of the North and South Pacific to the equatorial ocean are estimated in a coupled climate model. Idealized experiments show that the South Pacific contributes 30–50% more than the North Pacific to the equatorial ocean temperature change, because the warming in the North Pacific can induce an additional cold Rossby wave at around 20°N that tends to block the warm effect from the central North Pacific, while the South Pacific influence can be easily convoyed to the equator through the interior pathway. **Citation:** Yang, H., H. Jiang, and B. Tan (2005), Asymmetric impact of the North and South Pacific on the equator in a coupled climate model, *Geophys. Res. Lett.*, *32*, L05604, doi:10.1029/2004GL022195.

1. Introduction

[2] Previous studies have shown that the tropics can be modulated by the extratropics through the oceanic tunnel [*Gu and Philander*, 1997; *Wang and Weisberg*, 1998; *Kleeman et al.*, 1999]. On the one hand, the extratropical signal can be conveyed equatorward by the mean subduction flow [*Schneider et al.*, 1999] or planetary wave [*Stephens et al.*, 2001] through the interior pathway, and (or) further by the lower latitude western boundary current [*McCreary and Lu*, 1994; *Huang and Liu*, 1999] and Kelvin waves [*Lysne et al.*, 1997]. These processes are closely related to the Pacific shallow meridional overturning circulation – the subtropical cell (STC) [*McCreary and Lu*, 1994], and can be summarized as the mean advection mechanism. On the other hand, the extratropical anomalies can induce changes in overlaying atmospheric circulations that could eventually result in changes in the STC strength. The tropical climate can be thus changed due to the varying amount of the equatorward cold water transport [*Kleeman et al.*, 1999; *Nonaka et al.*, 2002], which is referred as the perturbation advection mechanism.

[3] The quantitative impact of the extratropics on the equator has been assessed in ocean general circulation models (OGCM) [*Shin and Liu*, 2000; *Nonaka et al.*, 2002; *Yang et al.*, 2004, hereinafter referred to as YLW04] and in a fully coupled climate model [*Liu and Yang*, 2003, hereinafter referred to as LY03]. OGCM studies suggest that the extratropical thermal forcing of 30° poleward [*Shin and Liu*, 2000] and the off-equatorial wind forcing of 10–25° [*Nonaka et al.*, 2002] are very efficient to disturb the equatorial thermocline through the mean advection mechanism and the perturbation advection mechanism. Furthermore, OGCM also shows that the impact of the South Pacific on the equatorial thermocline is much stronger than that of the North Pacific (YLW04). Coupled model study has shown a significant extratropical control of tropical climate in both the SST and the thermocline (LY03). However, it still remains unclear that the relative contributions of the South and North Pacific on the tropical climate in a fully coupled system.

[4] Using the same coupled climate model and modeling technique as by LY03, we further perform four partial coupling experiments to explicitly quantify the relative contributions of the Northern Hemisphere (NH) and Southern Hemisphere (SH) to the equator. The model is the Fast Ocean-Atmosphere Model (FOAM) [*Jacob*, 1997] that consists of the atmospheric component (R15) of NCAR-CCM3 and the ocean component of GFDL-MOM. Similar to experiments ABOT (Atmosphere Bridge/Ocean Tunnel) and OT (Ocean Tunnel) by LY03, four partial coupling experiments (N-ABOT, S-ABOT, N-OT and S-OT) are integrated for 200 years. In N-ABOT, a homogenous 2°C SST warming is "seen" by both the atmosphere and ocean only in the global NH (north of 30°N), and the mean SST seasonal cycle of the control run (CTRL) is prescribed in the global SH (south of 30°S). Instead, in N-OT, only the ocean "sees" the 2°C SST warming and therefore contributes to the southward subduction of extratropical SST anomaly. The configurations in S-ABOT and S-OT are opposite to those in N-ABOT and N-OT, respectively. The atmosphere and ocean between 30°S and 30°N are still fully coupled in these four experiments. Each experiment has its own CTRL run that is performed similarly as each experiment except without the 2°C SST warming. Our results show that the asymmetric impact of the NH and SH oceans on the tropical climate is also presented in this coupled model as that in the OGCM (YLW04). The climate change in the SH plays a dominant role not only in the equatorial mean climate changes in both the SST and the thermocline, but also in modulating the interannual variability such like El Nino-Southern Oscillation (ENSO).

2. Asymmetric Impact on Tropical Climate

[5] Through both the atmospheric bridge and oceanic tunnel (N-ABOT and S-ABOT), the SH ocean contributes 30–50% more than the NH ocean to the equatorial SST and thermocline, respectively. A 2°C NH SST warming can warm the equatorial SST by about 0.40°C (solid line, Figure 1a), while the same SH warming can cause the equatorial SST rise by 0.53°C (dashed line). Through only the oceanic tunnel (N-OT and S-OT), the equatorial SST changes in S-OT and N-OT are almost the same (dotted and dot-dashed lines, Figure 1a), however, equatorial thermocline temperature

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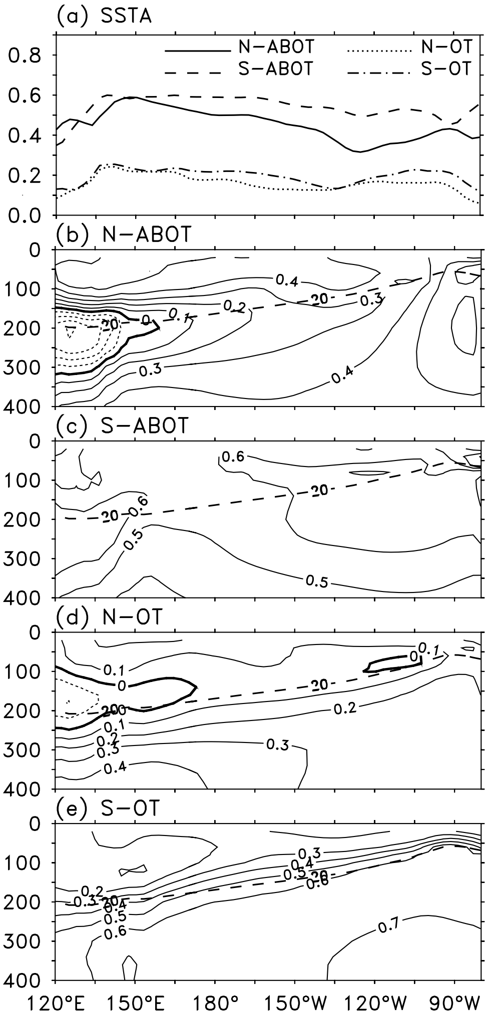

Figure 1. Mean climate changes in (a) SST and thermocline temperature for (b) N-ABOT, (c) S-ABOT, (d) N-OT and (e) S-OT averaged in a 5°N–5°S equatorial strip of the Pacific for model year 150–200. In (a) the solid line is for N-ABOT, dashed line for S-ABOT, dotted line for N-OT and dash-dotted line for S-OT. The depth of the 20°C isotherm from control run is plotted as long-dashed line in (b)–(e). The temperature change in each experiment is relative to its own control run, which has the identical setting as the respective partial coupling experiment except for the absence of SST anomaly on the prescribed seasonal cycle of control.

change in S-OT is much larger than that in N-OT (Figures 1d and 1e). These explicitly suggest a stronger SH contribution to the equatorial climate.

[6] The larger contribution to the equatorial SST and thermocline from the South Pacific can be attributed to the larger atmospheric Hadley cell change and therefore the larger STC change in the SH (LY03). In ABOT of LY03, the 2°C SST forcing is added in both the hemispheres so that the impacts of the NH and SH oceans on the equator are mixed together. Even though, we can still clearly see the much stronger equatorward warm anomaly subduction from the South Pacific as well as the larger SH Hadley cell change (figure not shown here). LY03 also stated that the larger South Pacific contribution can be expected because the north influence appears to be blocked by the ITCZ wind forcing and further weakened by the Indonesian Throughflow [*Rodgers et al.*, 1999].

[7] The four experiments performed in this work reveal that another mechanism, the Rossby wave originated at around 20°N of the eastern boundary of the Pacific, may significantly weaken the North Pacific influence. One consequence of the Rossby wave is to generate a cooling anomaly in the thermocline of the western equatorial Pacific (Figure 1b). This cooling anomaly is resulted from the NH anomalous forcing in N-ABOT and does not appear in S-ABOT (Figure 1c). It is located at between 150–300 m depth of the western equatorial Pacific thermocline, with a magnitude of about −0.3°C. It is clearly seen that this cooling anomaly tends to propagate eastward along the equatorial thermocline (defined as the depth of 20°C isotherm) as a cold equatorial Kelvin wave (Figure 1b), and outcrops to the surface in the eastern equatorial Pacific. This eventually results in a less warming in the SST there (Figure 1a), even though the subsurface temperature warms more in the east (Figure 1b). The maximum west-east SST contrast in N-ABOT (Figure 1a) reaches 0.3°C, which is considerably large given the mean SST change of 0.4°C. In contrast, the equatorial SST is nearly uniformly warmed in S-ABOT.

[8] The horizontal temperature changes in isopycnal layer averaged between 24 and 26 σ_t (Figure 2) explicitly show that the cooling anomaly in N-ABOT comes from the eastern boundary of the off-equatorial Pacific that is related to the westward baroclinic Rossby wave [*Liu*, 1999] or so-called westward subtropical pathway [*Zhang and Liu*, 1999]. The westward propagation of the cold Rossby wave manifests itself clearly in N-ABOT and appears to block most of the warm anomaly subduction originated in the North Pacific ventilation zone (Figure 2a). This cold Rossby wave is generated by a positive Ekman pumping anomaly at around 20°N along the western coastal of North America (Figure 3). The Ekman pumping velocity w_e is calculated as $curl(\boldsymbol{\tau}/f)/\rho$, where $\boldsymbol{\tau} = (\tau^x, \tau^y)$ is wind stress vector, f is Coriolis parameter and ρ is the sea water density. The Ekman pumping anomaly is derived as the difference of w_e between N-ABOT and its control. This positive Ekman pumping anomaly over the eastern subtropical Pacific can cause an upward displacement of the eastern boundary thermocline and thus a Rossby wave propagating westward

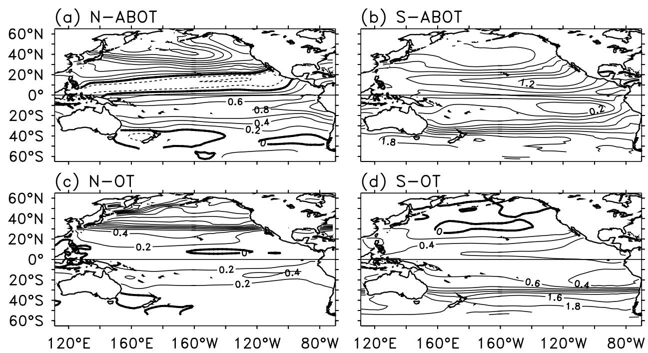

Figure 2. Horizontal patterns of the mean temperature changes averaged in year 150–200 between 24 and 26 σ_t for (a) N-ABOT, (b) S-ABOT, (c) N-OT and (d) S-OT. The contour interval is 0.2°C.

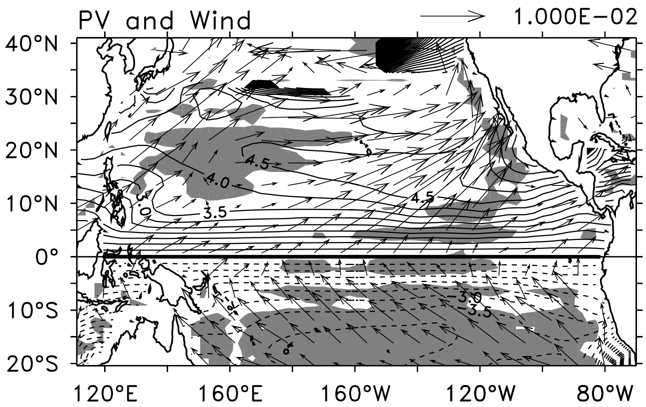

Figure 3. Mean potential vorticity (unit: $1 \times 10^{-10}\mathrm{m}^{-1}\mathrm{s}^{-1}$; contours) between 24 and 26 σ_t in control run, the surface wind stress change (unit: 10^{-2} N/m^2; vectors) and the change in Ekman pumping velocity w_e (shaded region) in N-ABOT. Only the upward Ekman pumping velocity change is plotted as the shaded region. See color version of this figure in the HTML.

with cold temperature anomaly [*Liu*, 1999]. The positive w_e anomaly is resulted from both the wind curl term $curl(\boldsymbol{\tau})/(\rho f)$ and the divergent of the surface Ekman flow $\beta\tau^x/(\rho f^2)$. Because of the given strong anomalous warming poleward of 30°N in N-ABOT, there is strong anomalous convergent surface air flow towards the higher latitudes of the NH (Figure 3), which induces an anomalous positive wind curl off the coast of the North America, and finally results in the positive Ekman pumping anomaly there.

[9] The cold Rossby wave tends to propagate westward along the lines of potential vorticity (PV) [*Stephens et al.*, 2001]. The mean PV field between 24 and 26 σ_t is calculated as $f\Delta\rho/\rho\Delta z$, where density difference $\Delta\rho = 26 - 24 = 2$ kg/m^3, Δz is the thickness between 24 and 26 isopycnal level. The alignment of the negative temperature anomaly in Figure 2a is in agreement with that of the maximum PV tongue in Figure 3, suggesting a PV control of the Rossby wave propagation.

[10] The negative temperature anomaly in the off-equatorial Pacific thermocline, although happened in the idealized global NH homogeneous warming experiment, has its counterpart in reality. Observational studies of the Pacific decadal and interdecadal variability have shown that an anomalous warming in the central North Pacific is usually accompanied by an anomalous cooling along the west coast of North America and vice versa [e.g., *Deser et al.*, 1996; *White and Cayan*, 1998]. The eastern boundary cooling is associated with coastal upwelling process. OGCM experiments forced by observed winds and SST have suggested that the anomalous cooling tend to block and cancel in part the subducting warm anomaly that originated in the central North Pacific, and finally results in a less than expected variability in the equator (YLW04). The coupled experiment N-ABOT suggests that the observed eastern boundary cooling might be closely related to the anomalous warming in the central North Pacific, and it can be thought as a result of the dynamic adjustment of the tropical coupled system to the extratropical forcing.

[11] The North Pacific negative thermocline anomaly in Figure 2a does not have counterpart in the South Pacific due to the different land-sea contrast and the asymmetric PV field in the two hemispheres (Figure 3). This explains the much stronger response of the tropical coupled system to the SH forcing as shown in Figures 1 and 2. Through the oceanic tunnel, the SH forcing can efficiently disturb the equatorial SST and subsurface temperature by both the mean advection mechanism and perturbation advection mechanism (Figures 1c, 1e, 2b, and 2d). It is worthy noting that the South Pacific anomalous warming can be easily subducted to the equator through the interior route without any blocking from the mean current, and warm the central-eastern Pacific thermocline first (Figures 1c and 1e). The impact of the South Pacific can cross the equator and further disturb the North Pacific thermocline (Figures 2b and 2d). A dynamic connection likely exists between thermocline variabilities in both hemispheres by the eastward equatorial Kelvin waves and westward off-equatorial Rossby waves excited along the western coast of America (YLW04). Therefore, the anomalous signal in the eastern North Pacific in Figures 2b and 2d may come from the eastern equatorial Pacific by means of coastal Kelvin waves, which may be excited by the eastward equatorial Kelvin waves.

[12] The stronger SH contribution to the equatorial mean climate suggests that it may also play a dominant role in modulating ENSO variability. The ENSO behavior in a warm climate could change due to changes in the equatorial trade winds, the vertical temperature stratification and the meridional overturning circulation, as the studies by *Timmermann et al.* [1999] and *Yang et al.* [2005]. Experiments N-ABOT and S-ABOT reveal that the 20% amplitude reduction in ENSO variability in ABOT of *Yang et al.* [2005] is, as expected, caused predominately by the South Pacific warming (figure not shown) because of the higher efficiency of the South Pacific to change the tropical coupled system (Figure 1).

3. Summaries

[13] As a complementary study of LY03 and *Yang et al.* [2005], this work explicitly shows the dominant role of the South Pacific in the modulation of the tropical climate. Besides the negative effects of the ITCZ wind forcing and the Indonesian Throughflow on the North Pacific influence, the North Pacific warming can induce an additional cold Rossby wave at around 20°N that block and cancel the warm effect from the central North Pacific. While the South Pacific influence can be easily convoyed to the equator without any blocking by the mean current and can further disturb the off-equatorial North Pacific. As a result, the South Pacific also plays more important role in the modulation of ENSO variability.

[14] The dominant influence of the SH ocean on the global climate can be speculated because the sea surface area in the SH is about 37% greater than in the NH. The larger contribution of the South Pacific is consistent with observations [*Johnson and McPhaden*, 1999] and may also shed light on paleoclimate records that the tropical temperature evolves synchronously with the Antarctic air temperature and atmospheric CO_2, but leads the NH continental ice volume [*Lea et al.*, 2000].

[15] This work shows an asymmetric impact of the hemispheric forcing on the equator in a fully coupled

model. The robustness of our estimate, however, needs to be tested in other models. Furthermore, more realistic forcings are needed to assess the role of the interactions in the observed climate change and climate variability.

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