Response of the Indo-Pacific warm pool to interannual variations in net atmospheric freshwater

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[1] We used the ocean general circulation model (OGCM) developed in the Massachusetts Institute of Technology to study the response of the Indo-Pacific warm pool (IPWP) to net atmospheric freshwater flux (NAFW, equal to evaporation minus precipitation) at interannual timescales. The OGCM is forced by observed, monthly NAFW from 1988 to 2000. Our simulations show that the magnitude of interannual anomalies of salinity and temperature reaches about 0.7 practical salinity unit and 0.4°C, respectively. The typical timescale of these interannual variabilities is about 3–5 years. Averaged over 1988–2000, a decrease (increase) of temperature is accompanied with a decrease (increase) of salinity in the western Pacific and South Indian Ocean. A decrease of temperature, however, is accompanied with an increase of salinity in the surface layer (0–50 m) of the North Indian Ocean. The diagnosed budgets of salinity and temperature (heat) are analyzed to estimate the role of advection and vertical mixing in response to the surface NAFW forcing. The analyses indicate that the salinity anomaly in the IPWP is largely due to vertical mixing, especially in the surface layer. The vertical mixing of salinity, in turn, is associated with the surface NAFW anomaly. In contrast, the temperature anomaly above 300 m is primarily due to changes in advection forced by the NAFW, which is associated with basin-wide changes in major ocean currents. Because of the strong effect of advection on the interannual variability of temperature, the temperature anomaly in the surface layer lags the salinity anomaly about 14–15 months. The results of our simulations are consistent with previous studies about the nearly immediate response of the tropical upper ocean to NAFW forcing due to vertical mixing. The slower response due to changes in basin-scale heat advection suggests the possibility that ocean variability at interannual and longer timescales can be generated by large-scale NAFW forcing at seasonal and longer timescales.

INDEX TERMS: 4215 Oceanography: General: Climate and interannual variability (3309); 4255 Oceanography: General: Numerical modeling; 4283 Oceanography: General: Water masses; 4504 Oceanography: Physical: Air/sea interactions (0312); 4512 Oceanography: Physical: Currents; KEYWORDS: Indo-Pacific warm pool, net atmospheric freshwater, interannual variability


1. Introduction

[2] It is well known that the surface water from the tropical eastern Indian Ocean to the tropical western Pacific Ocean is the warmest (approximately 28°C) on the Earth, which is frequently referred to as the Indo-Pacific warm pool. The Indo-Pacific warm pool (IPWP) is the major region of atmospheric convection and the major source of heat for driving global atmospheric circulation [see, e.g., Gadgil et al., 1984]. Since the relationship between SST and saturation vapor pressure via the Clausius-Clapeyron equation is nonlinear, there is a dramatic increase in atmospheric moisture content and atmospheric convection when the underlying SST crosses the threshold of approximately 28.5°C at tropical latitudes. Studies on the interannual variabilities of SSS and SST are mostly focused on the El Niño events [Philander, 1990], using available observations in the Pacific Ocean [Delcroix and McPhaden, 2002; Matsuura and Iizuka, 2000; Delcroix et al., 2000a].

[3] Using the 28.5°C threshold SST as the criterion for defining the surface area of the IPWP, V. M. Mehta and A. V. Mehta (manuscript in preparation, 2004) have shown that the IPWP area has undergone variability at interannual to multidecadal timescales in the twentieth century (1909–1998) historical SST data. Analyses of the NCEP-NCAR model-assimilated atmosphere data (1949–1998) showed (Mehta and Mehta, manuscript in preparation, 2004) that there is a strong association between the IPWP area, highest SST, and large-scale atmospheric convergence/divergence patterns.

[4] The rainfall associated with atmospheric convection over the IPWP can cause the formation of barrier layers in the underlying ocean [Lindstrom et al., 1987], which is a
topic of active observation and modeling studies [Godfrey and Lindstrom, 1989; Lukas and Lindstrom, 1991; Delcroix et al., 1992; Sprott and Tomczak, 1992; You, 1995]. These studies found that the barrier layers exhibit strong variability at various timescales including their complete erosion [Sprott and McPhaden, 1994; Roemmich et al., 1994; Vialard and Delecluse, 1998b]. Observations also showed [Feng et al., 2000] that rainfall is important in the upper ocean salinity budget and the formation and erosion of barrier layers. Observations and model results [Anderson et al., 1996; Feng et al., 2000; Schneider and Barnett, 1995] showed that the exchange of freshwater flux between the ocean and atmosphere is as important as the exchange of heat to the upper ocean buoyancy budget in the IPWP. Freshwater from rainfall, however, mixes vertically only during strong wind events that are usually associated with intraseasonal variability [Cronin and McPhaden, 2002]. Vialard and Delecluse [1998a, 1998b] found that the simulated SSS structure is sensitive to atmospheric freshwater, and the simulated barrier layer displays a strong interannual variability in the WPWP region. These results suggest that coupled atmosphere-ocean processes may be responsible for interannual-multidecadal variability of the IPWP, in which interactions between upper ocean and the atmosphere via the net atmospheric freshwater flux (NAFW), defined as evaporation minus precipitation at the surface, may be very important.

With a series of experiments using a model of the tropical-subtropical Pacific Ocean, Yang et al. [1999] showed that a reduction in NAFW decreases salinity as expected and increases temperature of the upper ocean. It was found that NAFW estimates based on microwave sounding unit radiances immediately (in the same season) result in a 0.6°C warming in the western Pacific. It is not clear, however, how the IPWP responds to the NAFW and what the mechanisms of the IPWP temperature and salinity changes are. Therefore the present study was designed to address the following questions: How sensitive is the IPWP to remotely sensed precipitation and evaporation estimates? What is the IPWP response to interannual variability of the NAFW? What are the mechanisms of the IPWP response? These questions are addressed with a global ocean general circulation model (OGCM) forced by monthly evaporation and precipitation estimates for the 1988–2000 period. The OGCM and experiment design are described in section 2. Evaporation and precipitation data used in this study are described in section 3. The time-average response of the IPWP and its mechanisms are presented in sections 4 and 5. The interannual variability of the IPWP response and its mechanisms are presented in sections 6 and 7. The results are discussed in section 8.

2. MIT Ocean General Circulation Model and Experiment Design

The OGCM developed in the Massachusetts Institute of Technology [Marshall et al., 1997; Huang et al., 2003a, 2003b] was used in the present study. The model domain is global, with realistic topography. The latitudinal resolution is 0.4° near the equator, linearly increasing to 2° at and poleward of 20°S and 20°N. The longitudinal resolution is 2°. There are 30 levels in the vertical with a resolution of 10 m between the ocean surface and 50 m depth, 25 m between 50 and 200 m depth, and 50 m between 200 and 400 m depth. The horizontal diffusivity is 10^3 m^2 s^-1. The vertical diffusivity and viscosity are calculated from K-profile parameterization [Large et al., 1994]. The KPP scheme parameterizes nonlocal mixing, which requires solar radiation as a separate penetrative heating. The vertical diffusion and KPP mixing are calculated implicitly in the model.

The OGCM was “spun up” for 200 years from a motionless, initial state of annual average temperature and salinity from Levitus and Boyer [1994] and Levitus et al. [1994]. During this spin-up, the OGCM is forced by monthly wind stress climatology from Helfer and Rosenstien [1983]. The first-layer salinity (S) is forced by satellite-based NAFW between 50°S and 60°N in 1988, and is restored to monthly SSS of Levitus and Boyer [1994] poleward of 55°S and 65°N with a timescale (τS) of 60 days as shown in equations (1a)–(1c):

\[
\frac{dS}{dt} = S_0(E - P)_{1988}/\Delta \zeta_1, \quad \text{45°S–55°N}, \quad (1a)
\]

\[
= (S SS - S)/\tau_S, \quad 65°-89°N \text{ and } 80°-55°S, \quad (1b)
\]

\[
= S_0(E - P)_{1988}/\Delta \zeta_1 + (S SS - S)/\tau_S, 45°-55°S \quad \text{and } 55°-65°N, \quad (1c)
\]

where \(\Delta \zeta_1\) is the first-layer thickness. A restoring surface boundary condition is applied poleward of 45°S and 55°N because the satellite-based NAFW is available only from 45°S to 55°N.

The first-layer temperature is forced by a mixed boundary condition, which consists of a restoration to the monthly SST of Levitus and Boyer [1994] with a 10-day damping timescale (\(\tau_T\)) and the monthly heat flux climatology from the Comprehensive Ocean Atmosphere Data Set [da Silva et al., 1994]:

\[
\frac{dT}{dt} = (Q_{\text{SOL}} + Q_{LS})/\rho C_P \Delta \zeta_1 + (SST - T)/\tau_T, \quad (2)
\]

where \(Q_{\text{SOL}}\) is the downward flux of solar radiation, and \(Q_{LS}\) represents the downward heat fluxes of latent, sensible, and infrared radiation. The mixed surface boundary condition of equation (2) can simulate well both SST and net heat flux into the ocean. While the addition of freshwater to the ocean implies the addition of heat, it is assumed here that the freshwater and the sea surface are at the same temperature. At the end of year 200, upper ocean salinity and temperature are reasonably close to the observed climatology in the equatorial Indo-Pacific Oceans as shown in Figure 1. Simulated average temperature between 5°S and 5°N is about 1°–3°C lower than observations in the eastern Pacific, but about 1°–3°C higher in the west. Simulated average salinity is about 0.5 practical salinity unit (psu) higher in the east, and 1.5 psu higher in the west. The larger errors of temperature and salinity in the western Pacific may partly be due to the mismatch of model and real topography. Simulated currents and Atlantic meridional overturning circulation (approximately 25 Sv at 40°N) are comparable to other OGCM simulations.
Figure 1. Observed (a) temperature and (b) salinity averaged between 5°S and 5°N from Levitus and Boyer [1994] and Levitus et al. [1994]. (c and d) Same as Figure 1a except for the simulation between 1988 and 2000. (e and f) Difference between the simulation and observation. Contour intervals (CIs) are 2°C in Figures 1a and 1c but are 1°C in Figure 1e. CIs are 0.25 psu in Figures 1b, 1d, and 1f. Negative values are dashed.
[9] After the OGCM was spun up, the net heat flux into the ocean is diagnosed as,

\[ Q = Q_{\text{Solar}} + Q_{LST} + \rho c_p \Delta T/\tau_f, \]

and the mixed surface boundary condition (2) is changed into the flux boundary condition

\[ dT/dt = Q/\rho c_p \Delta z_1 \]

in order to allow an SST perturbation to evolve according to ocean thermodynamics and dynamics. Control (CTR) and perturbation (PTB) experiments are then started from this state. The boundary conditions (1) and (4) are applied in the CTR experiment, in which the monthly NAFW of 1988 is used. The same boundary conditions are applied in PTB experiment except that the monthly NAFW from 1988 to 2000 is used. Both CTR and PTB experiments are integrated for 13 years from 1988 to 2000. The difference between the PTB and CTR runs is defined as the anomaly in this paper except otherwise specified. The use of the 1988 forcing field in the CTR run is to ensure that there is a smooth evolution of the PTB run from 1988 to 2000.

3. Satellite-Based Precipitation and Evaporation Estimates

[10] Monthly precipitation estimates used in this study are from the GPCP [Huffman et al., 1997] from January 1988 to December 2000. Monthly evaporation estimates used in this study are based on SSM/I [Chou et al., 1997] radiance for the same period. The SSM/I-based evaporation estimates were validated by in situ measurements in the WPWP region [Chou et al., 1997; Atlas et al., 1996], so they are well suited for the present study. The SSM/I-based evaporation estimates are available on a 1° × 1° grid over the oceans from 50°S to 60°N, whereas the GPCP precipitation estimates are available on a 2.5° × 2.5° grid from 90°S to 90°N. Monthly evaporation and precipitation estimates are interpolated to the OGCM grid and monthly NAFW is calculated on the OGCM grid.

[11] Analyses of NAFW estimates between 1988 and 2000 show that precipitation exceeds evaporation by approximately 100–150 cm/yr in the IPWP in the SPCZ and ITCZ regions (Figure 2a). Evaporation exceeds precipitation by approximately 150–200 cm/yr in the western (west of 60°E) and southern Indian Ocean, and the subtropics of the North Pacific and eastern South Pacific. The NAFW anomaly (Figure 2b) is approximately 25–50 cm/yr in the IPWP, and 100 cm/yr in the southwestern Pacific. In contrast, the NAFW anomaly is about −75 cm/yr near ITCZ.


[12] As indicated in Figure 2b, an average anomaly of NAFW (25–50 cm yr⁻¹) is found in the IPWP. Therefore, we first look into the average response of the IPWP to the NAFW anomaly. Vertical structures of the average (1988–2000) salinity and temperature differences between the PTB and CTR experiments are shown in Figures 3a and 3b, respectively. The salinity and temperature anomalies are confined to the upper 150–200 m (above 20°C isotherm) during the simulation period. In response to the increased NAFW in the WPWP (Figure 2b), the average salinity between 10°S and 10°N increases by approximately 0.2–0.5 psu in the WPWP above 100 m. Salinity decreases by approximately 0.2 psu near 180° longitude above 150 m in response to the decreased NAFW (Figure 2b). The average temperature increases by approximately 0.2°C in the western Indian Ocean between 50 and 150 m, by approximately 0.4°C in the EIPWP between 50 and 100 m, and by approximately 0.2°C in the WPWP above 150 m. In contrast, the temperature decreases by approximately 0.3°C in the western Pacific between 150°E and 180° above 200 m.

[13] To evaluate the horizontal structures of salinity and temperature anomalies, the upper ocean is divided into two layers based on the relative magnitude of temperature anomalies: a surface layer between 0 and 50 m and a subsurface layer between 50 and 150 m, because the horizontal structures of salinity and temperature anomalies within these two layers are very similar. In the surface layer (Figure 4a), salinity increases by approximately 0.2–0.5 psu in the EIPWP and southwestern Pacific. It decreases by approximately 0.3–0.4 psu in the western Pacific north of 15°S. In the subsurface layer (Figure 4b), the increase of salinity in the EIPWP is smaller (0.1–0.2 psu) than that in the surface layer. The subsurface salinity decreases by 0.1–0.3 psu in western Pacific between 10°S and 20°N.

[14] Compared to the horizontal structures of salinity anomalies, the horizontal structure of temperature anomalies is more complex. In the surface layer (Figure 5a), temperature decreases by approximately 0.2°C in the North Indian Ocean and western Pacific east of 150°E. Temperature increases by approximately 0.2°C in the WPWP and the Indian Ocean west of Australia. In the subsurface layer (Figure 5b), the cooling in the western North Indian Ocean is about 0.2°C, but it is surrounded by a 0.1–0.4°C temperature increase in the EIPWP and southern Indian Ocean. In the WPWP, temperature increases approximately 0.1°–0.2°C, but decreases about 0.4°C east of 150°E between 10°S and 15°N.

[15] The large-scale anomalies of salinity and temperature due to NAFW generate changes in the surface and subsurface circulations in the IPWP regions (not shown). The SEC and EUC increase approximately 1 cm s⁻¹, but the NECC decreases by 0.5 cm s⁻¹, although the horizontal and vertical structures of these major currents do not change very much. The equatorial divergent flow in the surface and equatorward convergent flow increases approximately 0.2 cm s⁻¹ in the WPWP, so that the tropical cell [Lu et al., 1998] becomes stronger. As described later, these circulation changes cause substantial changes in salt and heat advection in the IPWP.


[16] To study mechanisms of salinity and temperature anomalies between PTB and CTR, the terms governing
local changes of salinity and temperature are diagnosed as follows:

$$\partial_t (S, T) = (S, T)_{\text{Adv}} + (S, T)_{\text{Mix}},$$

where $\partial_t (S, T)$ are local changes of salt and temperature, and

$$(S, T)_{\text{Adv}} = -u \partial_x (S, T) - v \partial_y (S, T) - w \partial_z (S, T),$$

are salt and temperature advections, and

$$S_{\text{Mix}} = \partial_z S + S_{\text{KPP}} + S_0 (E - P)/\Delta z_1,$$

$$T_{\text{Mix}} = \partial_z T + T_{\text{KPP}} + Q/(\rho c_p \Delta z_1),$$

represent the vertical mixing of salt and temperature. Here, $\partial_z (S, T)$ are vertical diffusion and $(S, T)_{\text{KPP}}$ are nonlocal diffusion due to KPP mixing, which are diagnosed explicitly. The combination of these terms is based on their similarity of dynamic property and the fact that the advection in three directions is largely cancelled, as well as the vertical and KPP mixing. The NAFW and nonsolar radiation heat flux ($Q$) are included in the vertical mixing as an upper boundary condition. The heat flux due to solar radiation is ignored since it is the same in both the PTB and CTR experiments. Other friction terms such as horizontal diffusion are important numerically to stabilize the simulation, but physically have little effect on salinity and temperature. Therefore they are ignored in this analysis. The budget terms are calculated at each time step during the experiments and their monthly averages are analyzed.

5.1. Salinity

[17] Figure 6a shows the diagnosed anomaly of local salinity change in the surface layer between 0 and 50 m. The anomaly of local salinity change is largely consistent with the simulated total salinity change as shown in Figure 4a. Differences are found near the western coast of Australia and near the Philippines. These differences may result from explicit diagnosis of vertical mixing in equations (7)–(8), since they are calculated implicitly during model integration. In most regions of the IPWP, the local salinity change appears
Figure 3. Simulated meridional average (10°S–10°N) of (a) salinity and (b) temperature anomaly between 1988 and 2000. CI is 0.1 psu and 0.1°C in Figures 3a and 3b, respectively. Thick contours are average temperature contours in °C.

to be associated with vertical mixing as shown in Figure 6c. Salinity anomalies due to vertical mixing in the IPWP are consistent with the local salinity change in Figure 6a, suggesting that the surface salinity anomaly is directly caused by vertical mixing, which, in turn, is associated with the NAFW at the surface. The salinity anomaly due to advection (Figure 6b) is approximately $100 \times 10^{-16}$ psu/s in the IPWP. Its sign is generally opposite to the local salinity change, except in the southern Indian Ocean where a positive salinity anomaly due to advection corresponds to a positive local salinity change of 5 to $10 \times 10^{-16}$ psu/s. This suggests that the local salinity change may have been caused largely by salt advection in this region.

[18] Similarly, in the subsurface layer between 50 and 150 m (not shown), anomalous local salinity change is consistent with simulated total salinity anomaly in Figure 4b. The local salinity change is mainly associated with the increase of salinity flux due to vertical mixing. The salinity flux due to advection is overall opposite to that due to vertical mixing and local salinity change.

5.2. Temperature

[19] Similar to the mechanisms of salinity change, the model diagnosis indicates that anomalous local temperature change is consistent with simulated total temperature anomaly. In contrast to the dominant effect of vertical mixing in salinity change, the temperature change due to advection dominates over that due to vertical mixing, especially in the subsurface layer as shown in Figure 7. The local temperature change is negative in the northern and equatorial Indian Ocean, and in the regions of Indonesian Throughflow and its outflow. However, it is positive to the west of 50°E and south of 13°S (Figure 7a). These local temperature changes appear to result from the temperature change due to advection (Figure 7b). Thus local temperature change is largely controlled by the heat flux due to advection above 150 m in
these OGCM experiments. However, vertical mixing may have a dominant effect on the local temperature change in the western Indian Ocean west of 60°E.

[29] Overall, the salinity anomalies were located near the surface (Figures 3a and 4), but the temperature anomaly seems to penetrate down into the subsurface (Figures 3b and 5). The association between NAFW and salinity anomalies is very clear, but it is unclear how the subsurface temperature anomaly is formed. Yang et al. [1999] indicated that the correlation between NAFW (SSS) and SST anomalies is negative during the precipitation event in the tropical Pacific in 1983, implying that the ocean temperature decreases (increases) when NAFW increases (decreases). Their analysis showed that the vertical mixing is increased when NAFW increases due to denser surface water. However, this mechanism does not seem to be valid in the long-term (13 years) average. Contrary to their conclusion, the present study indicates that the correlation between the anomalies of salinity and temperature is positive in the western Pacific (compare Figures 3a with 3b, Figures 4a with 5a). Therefore it is possible that the correlation between salinity and temperature anomalies is dependent on timescales. In addition, we should note that the structure of anomalous NAFW forcing in our simulation is much more complicated than those used by Yang et al. [1999]. Also, changes of salinity and temperature in our simulation may be forced not only by local anomalous NAFW, but also remote anomalous NAFW.


6.1. Large-Scale Variability

[21] Variabilities in monthly anomalies of NAFW, and upper layer salinity and temperature are shown in Figure 8. The standard deviation of NAFW is the largest (120 cm yr⁻¹)
in the South Indian Ocean near 8°S and 100°E and the WPWP near the equator and 170°E, as shown by Delcroix and McPhaden [2002]. Salinity variability is also the largest in the same two regions, with standard deviations approximately 0.4 psu, which is very close to the observation of 0.4 psu shown by Delcroix and McPhaden [2002]. Maxima of the standard deviations of lower layer salinity (not shown) are located in the same regions as in the upper layer, but are somewhat smaller. In contrast, the maxima of upper layer temperature variability are not colocated with those of NAFW or salinity maxima. They are located in the North Indian Ocean and the Indonesia archipelago. Maxima of the lower layer standard deviation are colocated with and have approximately the same magnitude of 0.2°C as in the upper layer, which is much smaller than the observation of 0.9°C from Delcroix and McPhaden [2002]. The reason is that the observed anomaly includes the effect of not only NAFW anomaly, but also the wind stress and surface heat flux anomalies. The cooling events in the eastern tropical Indian Ocean in 1994 and 1997 [Haugen et al., 2002; Huang and Kinter, 2002; Perigaud et al., 2003] are not well simulated, probably due to the lack of anomalous wind forcing.

[22] The vertical structures of EIWP and WPWP salinity anomalies are shown in Figure 9. The salinity anomalies in the surface layer (0–50 m) and the subsurface layer (50–150 m) have opposite signs. In contrast, temperature variations (Figure 10) in the EIWP and WPWP regions during the 13 years period occur largely at 50 to 250 m depths and do not appear to be forced directly by the surface NAFW anomalies. The vertical structures of salinity and temperature anomalies appear to be associated with the mechanisms
Figure 6. Simulated anomalies of salinity changes (1988–2000) of (a) local change, (b) advection, and (c) vertical mixing in the surface layer between 0 and 50 m. CIs are $5 \times 10^{-10}$ s$^{-1}$ in Figure 6a, but $50 \times 10^{-10}$ s$^{-1}$ in Figures 6b and 6c.
Figure 7. Simulated anomalies of temperature changes (1988–2000) of (a) local change, (b) advection, and (c) vertical mixing in the surface layer between 50 and 150 m. CIs are $5 \times 10^{-10}$ K s$^{-1}$ in Figure 7a, but $25 \times 10^{-10}$ K s$^{-1}$ in Figures 7b and 7c.
resulting in these anomalies. As shown in section 5, the salinity anomaly near the surface is largely due to surface NAFW, while advection is a major contribution in the subsurface ocean. The temperature anomaly is mainly determined by advection in both the surface and subsurface ocean. However, the effect of advection on temperature anomaly is much stronger in the subsurface ocean than in the surface ocean.

[23] Interannual variations in the NAFW, especially on the eastern edge of the WPWP, cause east-west variations of

Figure 8. Standard deviations of monthly (a) NAFW, (b) salinity, and (c) temperature in the PTB run between 1988 and 2000. CIs are 10 cm/yr, 0.05 psu, and 0.025°C, respectively, in Figures 8a, 8b, and 8c.
the salinity and temperature in the WPWP (Figures 11a, 11b, and 11d). Variations in the NAFW and salinity are also coherent in the EIWP. Precipitation moved eastward in the central and eastern equatorial Pacific during the El Niño events from mid-1991 to mid-1992 and from early 1997 to late 1998. These precipitation excursions can be seen from the stretching of negative NAFW anomalies in Figure 11a. The negative NAFW anomalies caused significant SST anomalies (0.5°C in late 1998) in the central and eastern equatorial Pacific, which spread westward (Figure 11d). This response is consistent with the conclusion of Yang et al. [1999] about the role of the NAFW in the 1982–83 El Niño event. SST variations in the IPWP, however, do not appear to be coherent with the surface NAFW variations, although significant east-west variations of SST are apparent in the IPWP.

[24] Compared with observations [Delcroix et al., 2000b] in Figure 11c, salinity anomaly is well simulated in both the magnitude and time evolution in the equatorial Pacific (Figure 11b). The indication is that the salinity anomaly is largely due to NAFW. However, the east-west migration of the salinity front in the WPWP is not as strong as in the observations from Matsukra and Iizuka [2000] during the El Niño event of 1991–1992. The reason might be the lack of strong westerly wind burst and associated eastward anomaly of the SEC in our simulation, as indicated in the observations from Delcroix et al. [2000a], Henin et al. [1998], Vialard and Delecluse [1998b], and Frankignoul et al. [1996]. The temperature anomaly is much smaller in the OGCM simulation (Figure 11d) than in the observations (Figure 11e). The magnitude of temperature anomaly in the OGCM simulation is 0.2°C and 0.5°C, respectively, in the

Figure 9. Hovmoller diagram of monthly salinity anomaly averaged in (a) EIWP and (b) WPWP in the PTB run. CIs are 0.05 psu. The area of warm pool is defined by SST ≥ 28.5°C.
IPWP and central Pacific. In contrast, they are about 0.5°C and 4°C in the observations. The phase of temperature anomaly is also different between the simulation and observations. These differences reflect the fact that the temperature anomaly results from the combined effect of interannual variations in wind stress, surface heat flux, and NAFW in the observations. In comparison, the temperature anomaly results only from the NAFW in the simulation. Nevertheless, the temperature anomaly due to the NAFW clearly contributes to the observed total temperature anomaly during the El Niño events in 1997–98 and 1991–92. In addition, the presence of decadal-scale anomalies of temperature is also evident in both the simulation and observations (Figures 10, 11d, and 11e).

[25] In the WPWP (region B, refer to Figure 5a), the salinity anomalies are well simulated (Figure 12a), since they are dominantly forced by surface NAFW via vertical mixing that will be discussed in the next subsection. For example, simulated salinity anomalies in the surface layer of WPWP are about 0.5 psu in 1992 and 1998 and −0.4 psu in 1989 and 1996. These anomalies are very consistent with observations in both magnitude and phase. However, simulated temperature anomaly is only 0.2°C, while the observed anomaly reaches 0.8°C (Figure 12b). The phase of simulated temperature anomaly is not coherent with observations either. This reflects the fact that the temperature anomaly is largely associated with surface wind stress and heat flux anomalies [Huang and Liu, 2001], while they are fixed in our simulation.

6.2. Regional Response of Salinity and Temperature

[26] As shown in section 4, the time-averaged anomaly of salinity and temperature reaches 0.5 psu and 0.4°C, respectively, in the IPWP region (refer to Figure 3). However,
what is their interannual variability? To answer this question, we chose two regions in the central EIWP (region A; refer to Figure 5a) and the central WPWP (region B) to represent the warm pool in the Indian Ocean and Pacific sectors. Each region is further divided into the surface (0–50 m) and subsurface (50–150 m) layers as explained in section 4.

[27] As described in section 5a, time-averaged salinity anomaly is largely controlled by the NAFW. Figure 13 shows the NAFW anomaly in regions A and B, along with its accumulation, which is defined as

\[ F = \int_{1988}^{t} (E - P) dt. \]  

(9)

The accumulated NAFW reaches 600 cm by year 2000, while NAFW anomaly fluctuates between −200 and
There is pronounced interannual-decadal variability of the NAFW. It is higher during 1997–98, but lower during 1995–96, and in 1999.

In region A between 0 and 50 m (Figure 14a), the salinity anomaly exhibits pronounced interannual variations: low during 1995–96, but high during 1991–94 and 1997–98, which are clearly associated with low-frequency NAFW variations (Figure 13a). The correlation coefficient between salinity and NAFW is about 0.48 (Table 1). However, the correlation coefficient between temperature and NAFW is almost zero (Table 2). Therefore there is no correlation between salinity and temperature variations (Table 3). The temperature decreases approximately by 0.2°C from 1988 to 2000. In the subsurface layer between 50 and 150 m (Figure 14c), the variation of salinity and temperature are correlated well, with a correlation coefficient of 0.87 (Table 3). They increase about 0.2 psu and 0.4°C, respectively, in 1993 and 1998.

In region B between 0 and 50 m (Figure 14b), the salinity and temperature exhibit a large anomaly during 1992–94 and 1998–99 with a magnitude of 0.7 psu and 0.3°C, respectively. The effect of seasonal cycle on temperature anomaly is clear in this region, although the seasonal cycle on salinity anomaly is not evident. The correlation
coefficients between salinity and NAFW, and between salinity and temperature are 0.48 and 0.30 (Tables 1 and 3), respectively. However, there is no correlation between temperature and NAFW. In the subsurface layer between 50 and 150 m (Figure 14d), salinity decreases 0.1 psu from 1988 to 1995, but increases back to the average from 1995 to 2000. Contrary to the evolution of salinity, temperature increases 0.3°C from 1988 to 1994, but decreases back to the average in the next three years, and increases 0.2°C in later years. The correlation coefficient between salinity and temperature is $-0.7$.

From the evolution of salinity and temperature, we found that their relationship is complex. Interannual variabilities of salinity and temperature are positively correlated in the surface layer of IPWP, but negatively correlated in the subsurface layer of WPWP. The variability at timescale of 3–5 years is evident, which may be associated with the ENSO signal in the NAFW forcing. The phase lag is relatively shorter (2–6 months, Table 1) between NAFW and salinity anomalies, but longer (17–18 months, Table 2) between NAFW and temperature anomalies due to slower adjustment of ocean currents and their impacts on temperature. As a result, the SST variability lags SSS variability about 14–15 months (Table 3). This will be discussed further in the next section. In addition, it appears that there are possible decadal variabilities of salinity and temperature in both the simulation and observations as indicated in Figure 11. However, this result is only suggestive because of the short simulation period (13 years).

7. Mechanisms of Interannual Response

As discussed in section 5, the average response of temperature to the NAFW anomaly is associated with advection, but the time-averaged response of salinity is largely due to vertical mixing. Important mechanisms which determine the interannual response of the IPWP to NAFW are now discussed in this section.

To assess the effects of advection and vertical mixing on the interannual variability of salinity and temperature, we calculate the variations of salinity and temperature due to

**Figure 12.** Anomalous (a) salinity and (b) temperature from simulation and observation in Region B. Anomalies are from monthly data except for observed salinity, which is annual data retrieved from http://www.nodc.noaa.gov.

**Figure 13.** Observed NAFW anomaly (left coordinate, cm/yr) and accumulated freshwater flux (F, right coordinate, cm) in (a) region A and (b) region B.
advection and vertical mixing separately. The calculation is based on the model diagnosis conducted at each time step during the simulation from 1988 to 2000 (equations (5)–(8)). The total (local) salinity and temperature anomalies are separated into anomalies due to advection and vertical mixing as follows:

\[(\Delta S, \Delta T)_{\text{Total}} = \sum_{1988}^{t} (S_i, T_i) \cdot \Delta t \] (10)

\[(\Delta S, \Delta T)_{\text{Adv}} = \sum_{1988}^{t} [-u \Delta \lambda_i(S, T) - v \Delta \lambda_i(S, T) - w \Delta \theta_i(S, T)] \cdot \Delta t, \] (11)

\[(\Delta S, \Delta T)_{\text{Mix}} = \sum_{1988}^{t} (S_{\text{Mix}}, T_{\text{Mix}}) \cdot \Delta t, \] (12)

where \(\Delta\) represents the terms on the right-hand side of equations (10)–(12) are anomalies between the PTB and CTR runs. \(S, T, S_{\text{Mix}}, T_{\text{Mix}}\) and \(\Delta t\) are from equations (5)–(8). \(\Delta t\) is the time interval of one month between model diagnosis outputs averaged every month from the terms calculated at each time step. By comparing the sign and tendency of anomalies due to advection and vertical mixing with those of total anomalies, it is easy to see the relative importance of advection and vertical mixing in interannual variabilities. In the following, we will present these components in regions A and B (refer to Figure 5).

In region A of the EIW (Figure 15a), the variations of total salinity anomaly in the surface layer between 0 and 50 m is largely due to vertical mixing. The magnitude of salinity anomaly due to both advection and vertical mixing is large and almost linearly increasing during 1988–2000. In contrast, it is about 0.5 psu larger due to vertical mixing than due to advection. Therefore it appears that the total salinity anomaly is associated with vertical mixing in the surface layer. In the subsurface layer between 50 and 150 m (Figure 15c), however, the salinity anomaly of 0.2 psu due to advection becomes dominant.

### Table 1. Correlation Coefficient \(C_0\) Between Salinity and NAFW After Filtering Out Seasonal Cycle\(\text{a}\)

<table>
<thead>
<tr>
<th>Region</th>
<th>(C_0(S, \text{NAFW}))</th>
<th>(C_{\text{max}}(S, \text{NAFW}))</th>
<th>(L_S)</th>
</tr>
</thead>
<tbody>
<tr>
<td>A (0–50 m)</td>
<td>0.48</td>
<td>0.66</td>
<td>2</td>
</tr>
<tr>
<td>B (0–50 m)</td>
<td>0.48</td>
<td>0.84</td>
<td>6</td>
</tr>
</tbody>
</table>

\(\text{a}\) \(C_0\) is at 95% confidence level if it is greater than 0.17. \(C_{\text{max}}\) is the maximum correlation coefficient. \(L_S\) is the lag of salinity in months.

### Table 2. Same as Table 1, Except for Correlation Coefficient Between Temperature and NAFW\(\text{a}\)

<table>
<thead>
<tr>
<th>Region</th>
<th>(C(T, \text{NAFW}))</th>
<th>(C_{\text{max}}(T, \text{NAFW}))</th>
<th>(L_T)</th>
</tr>
</thead>
<tbody>
<tr>
<td>A (0–50 m)</td>
<td>−0.01</td>
<td>0.25</td>
<td>17</td>
</tr>
<tr>
<td>B (0–50 m)</td>
<td>−0.06</td>
<td>0.47</td>
<td>18</td>
</tr>
</tbody>
</table>

\(\text{a}\) \(L_T\) is the lag of temperature in months.
Table 3. Same as Table 1, Except for Correlation Coefficient Between Temperature and Salinity

<table>
<thead>
<tr>
<th>Region</th>
<th>C(T, S)</th>
<th>C\text{max}(T, S)</th>
<th>L_f</th>
</tr>
</thead>
<tbody>
<tr>
<td>A (0–50 m)</td>
<td>0.00</td>
<td>0.48</td>
<td>15</td>
</tr>
<tr>
<td>A (50–150 m)</td>
<td>0.87</td>
<td>0.50</td>
<td>14</td>
</tr>
<tr>
<td>B (0–50 m)</td>
<td>0.30</td>
<td>0.74</td>
<td>14</td>
</tr>
<tr>
<td>B (50–150 m)</td>
<td>−0.71</td>
<td></td>
<td></td>
</tr>
</tbody>
</table>

*L_f is the lag of salinity in months.

The temperature anomaly in region A decreases due to advection in the surface layer between 0 and 50 m (Figure 16a), but increases due to vertical mixing from 1988 to 2000. However, it is slightly stronger due to advection than due to vertical mixing. The net effect is that the total temperature anomaly decreases by 0.2°C from 1988 to 2000, which is clearly due to advection. In the subsurface layer between 50 and 150 m (Figure 16c), the evolution and magnitude of total temperature anomaly almost follows that of the temperature anomaly due to advection, suggesting the dominant effect of advection on the interannual variability of subsurface temperature.

In region B of the WPWP between 0 and 50 m (Figure 15b), the total salinity anomaly is largely associated with vertical mixing in the WPWP. The salinity anomaly increases by 4 psu due to vertical mixing such as in region A, but decreases by about 3.5 psu due to advection. The total salinity anomaly of 0.5 psu is clearly due to vertical mixing. The relative maximum of total salinity anomaly in 1993 and 1999 are also consistent with those in the salinity anomaly due to vertical mixing. In the subsurface layer between 50 and 150 m (Figure 15d), the negative total salinity anomaly is largely due to advection before 1996, but is controlled by vertical mixing after 1996, although its overall tendency appears to be consistent with that of advection.

The total temperature anomaly in region B between 0 and 50 m (Figure 16b) increases to 0.3°C by 1994, decreases to zero by 1998, and increases slightly after 1998. The positive temperature anomaly is largely associated with vertical mixing, which increases by about 0.4°C by 2000. Therefore it appears that vertical mixing plays an important role in the total temperature anomaly in this region. However, the evolution of the total temperature anomaly and its peaks in 1994 and 1999 are clearly associated with advection in the subsurface layer between 50 and 150 m (Figure 16d). The temperature anomaly due to vertical mixing decreases monotonically to 0.5°C by 2000.

In brief, the mechanisms of interannual variability of salinity are largely consistent with those of the time-averaged salinity anomaly. Salinity variability is dominantly controlled by vertical mixing due to the NAFW forcing. The close correlation between salinity and NAFW forcing can be seen from their higher correlation coefficient up to

![Figure 15](image-url)
0.48 (Table 1) with salinity lagging approximately 2–6 months. The reason is that the salinity anomaly is directly associated with local vertical mixing. In contrast, the impact of the NAFW on the temperature is first to change currents and then heat advection. The adjustment of currents may originate from anomalous NAFW in remote regions. Therefore the temperature response lags the NAFW forcing longer (17–18 months, Table 2) with a correlation coefficient of 0.25–0.47. As a result, the temperature response in the surface layer lags the salinity response approximately 14–15 months with a maximum correlation coefficient of 0.48–0.74 (Table 3).

[38] In addition, we should note the seasonal fluctuations in anomalous temperature and salinity (Figures 14b and 14c), which appear to be associated with advection (Figures 15c and 16b). We should also note the large change in salinity and temperature before and after 1996. This may partly be associated with local NAFW anomaly (Figure 13), but it may also be due to nonlocal anomalous NAFW in the subtropics via advection.

8. Summary and Discussion

[39] We studied IPWP responses to satellite-based NAFW estimates from 1988 to 2000, using the MIT OGCM. The time-averaged anomalies of salinity and temperature due to the NAFW forcing between 1988 and 2000 are approximately 0.5 psu and 0.4°C, respectively, in the northeast Indian Ocean. The amplitudes of salinity and temperature anomalies in the interannual timescale reach approximately 0.7 psu and 0.4°C, respectively, in the central EIWP and WPWP. Simulated salinity anomaly due to NAFW is consistent with observations in the WPWP and equatorial Pacific, indicating that the observed salinity anomaly is mainly associated with surface NAFW anomaly. The simulated temperature anomaly, however, is only about 0.2°C while the observed anomaly is approximately 0.8°C in the WPWP. The reason is that the simulated temperature anomaly is only due to the NAFW anomaly, while the observed temperature anomaly includes the effect from wind stress and surface heat flux. A similar problem is also found in the Indian Ocean, where exhibited major cooling events in 1994 and 1997 [Huang and Kinter, 2002; Haugen et al., 2002; Perigaud et al., 2003]. These cooling events are not simulated well, which may be due to the same reason that anomalous wind stress is not included in our simulation. As shown by Huang and Kinter [2002], these cooling events may largely be due to anomalous advection associated with anomalous wind stress.

[40] Analyses of mechanisms of salinity and temperature variations in the OGCM experiments indicate that the salinity anomalies are mainly due to vertical mixing, especially in the surface layer. The vertical mixing of salt, in
turn, is directly associated with the NAFW forcing. In contrast, temperature anomalies are largely associated with advection changes, especially in the subsurface layer.

[41] Effects of advection and vertical mixing on salinity and temperature are similar in both the time average and interannual responses. The temperature variability lags salinity variability by approximately 14–15 months in the surface layer of IPWP. On the basis of OGCM simulations and these analyses, we conclude that changes in horizontal, basin-wide heat advection play a major role in the interannual variability of IPWP temperature due to the NAFW forcing.

[42] As concluded by Yang et al. [1999], SST anomalies are negatively correlated with SSS anomalies in the central Pacific during an El Niño event in the central Pacific (Figures 11b and 11d). However, our study indicates that the time-averaged anomalies of salinity and temperature are positively correlated in the surface layer of IPWP (Figure 3). Their investigation showed that vertical mixing plays a dominant role in determining temperature anomalies due to the NAFW forcing at the seasonal timescale. The present study indicates that advection may play a dominant role in temperature anomalies in the interannual timescale, which suggests that the temperature in the surface and subsurface layers may be associated with remote NAFW forcing. To clarify the role of remote NAFW forcing, we designed an additional OGCM simulation forced by the NAFW anomaly in the IPWP region only. The results indicate (not shown) that the anomalies of temperature and salinity in the Indian Ocean remain almost the same as in the PTB run, suggesting that they are forced by the NAFW within the IPWP. However, the anomalies of temperature and salinity are changed significantly in the WPWP region, suggesting that they are largely associated with the NAFW away from the IPWP region.

[43] The role of ocean processes in the model may be dependent on model parameters and/or experiment design. Yang et al. [1999] applied a mixed boundary condition (as in equation (2)) during both spin-up and simulation periods, although the damping coefficient was reduced from 100 to 10 Wm$^{-2}$ K$^{-1}$. In contrast, we used a pure flux boundary condition (as in equation (4)). In the real world, the SST anomaly forced by NAFW may be damped by air-sea heat flux, but it may also be amplified due to feedback from air-sea freshwater flux. A method difference is that the vertical mixing parameterization by Pacanowski and Philander [1981] is applied by Yang et al. [1999], but the KPP mixing [Large et al., 1994] is used in the MIT OGCM.

[44] To assess the relative importance of various processes, we separated the total anomalies of salinity and temperature into anomalies due to advection and vertical mixing. Advection includes three different components. Vertical mixing includes vertical diffusion, nonlocal KPP mixing, and surface forcing. It would be very interesting to see the relative importance of various components of advection and vertical mixing in affecting the salinity and temperature anomalies. Our analysis showed that the NAFW plays a major role in the vertical mixing of salinity. For temperature advection, however, there is not one component that plays a dominant role in determining temperature anomalies. Different components may play important roles in different regions.

[45] A further question is what causes advection to change and play a dominant role in determining temperature anomalies. Our simulations indicate that changes in advection involve changes in basin-wide, Indo-Pacific ocean circulation. Therefore the changes in advection may not be necessarily driven by local NAFW forcing in the IPWP region. In our simulations, the major ocean currents in the tropical Pacific change about 1 cm s$^{-1}$. These current changes are consistent with enhanced NAFW in the IPWP and reduced NAFW in the central and eastern Pacific (Figure 1b), as indicated by Vialard and Delecluse [1998a] and Murtagudde and Busalacchi [1998]. For example, the cooling in the surface layer of the WPWP is caused by increased cold advection from the central equatorial Pacific due to a stronger SEC. The warming in the subsurface layer of the WPWP is associated with increased warm advection. The reason is that the subsurface convergence flow (warm advection) toward the equator increases due to positive subtropical NAFW forcing (Figure 1b), and the equatorial upwelling reduces due to positive NAFW forcing.

[46] Further studies are planned to compare the relative importance of wind stress, heat flux, and freshwater flux in affecting the IPWP and its variability.

Notation

CRCES Center for Research on the Changing Earth System;
CTR control run;
EIWP eastern Indian warm pool, SST > 28.5°C;
ENSO El Niño-Southern Oscillation;
EUC equatorial undercurrent;
GPCP Global Precipitation Climatology Project;
IPWP Indian-Pacific warm pool, SST > 28.5°C;
ITCZ Intertropical Convergence Zone;
KPP K-profile parameterization;
MIT Massachusetts Institute of Technology;
NAFW net atmospheric freshwater, evaporation minus precipitation;
NCAR National Center for Atmospheric Research;
NCEP National Centers for Environmental Prediction;
NEC north equatorial current;
NECC north equatorial countercurrent;
OGCM ocean general circulation model;
PTB perturbation run;
SEC south equatorial current;
SPCZ South Pacific convergence zone;
SSM/I Special Sensor Microwave/Imager;
SSS sea surface salinity;
SST sea surface temperature;
WPWP western Pacific warm pool, SST > 28.5°C.

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