Interannual-to-decadal variations of tropical-subtropical mass exchange in the Pacific Ocean: boundary versus interior pathways

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Abstract

Interannual-to-decadal variations of tropical-subtropical mass exchange in the Pacific Ocean are investigated using a near-global ocean general circulation model along with satellite observations of sea level and wind and a data assimilation product. The analysis focuses on the variability of pycnocline transports through the western boundary and interior near 10°N and 10°S. In contrast to time-mean exchange where boundary and interior pycnocline transports are both equatorward, the variations of boundary and interior pycnocline transports are found to be generally anti-correlated to each other. Moreover, the variation of boundary pycnocline transport is smaller than that of the interior, again different from time-mean exchange where boundary transport at 10°N is substantially larger than that through the interior. Interannual variation of boundary and interior transports are consistent with near-surface geostrophic flow inferred from sea level data. Interior pycnocline flow into the tropics is weaker in the 1990s than that in the 1980s, in agreement with recent observation. However, approximately half of it is compensated by opposite change in boundary flow at 10°N. The results indicate that the interior pathway is more important to interannual and decadal variability of tropical-subtropical exchange than the boundary pathway, despite a much larger time-mean transport of the western boundary current at 10°N. To a large extent, the counteracting tendency of boundary and interior flow and larger variation of the latter can be explained by the combined effect of variability in off-equatorial wind stress curl in the western Pacific and near-equatorial zonal wind stress. The former changes the strength of horizontal circulation and results in variation of boundary pycnocline flow that is opposite in direction but comparable in magnitude to that of the interior pycnocline flow. The latter primarily affects the strength of the shallow meridional overturning circulation with net pycnocline flow (mostly in the interior) opposing the surface Ekman flow. The co-variability of these two forcings leads to an enhancement of interior transport. The relative variability of boundary and interior pycnocline flow is insensitive to whether the Indonesian throughflow is present or not.
1. Introduction

Water mass exchange between the tropical and subtropical Pacific Oceans plays an important role in the maintenance of pycnocline property in the tropical Pacific Ocean and is believed to affect interannual-to-decadal variability in the tropics (Gu and Philander 1997, Zhang et al. 1999, Federov and Philander 2001, McPhaden and Zhang 2002). The exchange is accomplished by the so-called subtropical cell or STC (McCreary and Lu 1994) that connects the upper tropical and subtropical Pacific Oceans. Observational and modeling studies have advanced the knowledge about the pathways, strength, and mechanisms through which subtropical waters intrude the Equatorial Pacific Ocean (e.g., Fine et al. 1987, McPhaden and Fine 1988, McCreary and Lu 1994, Liu et al. 1994, Liu and Philander 1995, Lu and McCreary 1995, Rothstein et al. 1998, Huang and Liu 1999, Zhang et al. 1999, Harper 2000, Huang and Wang 2002). A relatively common picture emerges from these studies: waters subducted in the eastern part of the northern and southern subtropical gyres (if not too far east) travel westward and equatorward to the tropical Pacific Ocean where they join the equatorial undercurrent either through low-latitude western boundary currents (LLWBCs) or through the interior ocean. These waters upwell near the equator and then return to the subtropical gyres as Ekman transport. It has been established that pycnocline waters from the subtropics arrive at the tropics via two pathways: through the western boundary via the LLWBCs and through the interior.

The latitudes near 10°N and 10°S are critical to tropical-subtropical exchange because it is there that pycnocline waters carried by the North Equatorial Current (NEC) and South Equatorial Current (SEC) bifurcate into mid- and low-latitude western boundary currents: the NEC splits into the Kuroshio and the Mindanao Current (MC) whereas the SEC branches into the East Australian Current and the New Guinea Coastal Undercurrent (NGCU). In fact, a hydrographic section was occupied at 10°N as part of the World Ocean Circulation Experiment to understand exchange between the tropical and northern subtropical Pacific Ocean. Many studies suggest that mean boundary pycnocline transport is substantially larger than that through the interior near 10°N, but the difference is smaller near 10°S (cf. summary by Liu and Philander 2000).
However, less is known about the variability of boundary and interior exchange on interannual and decadal time scales that is potentially important to tropical heat content variability. Lukas (1988) found that the transport of the MC inferred from tide-gauge data was weaker and stronger before and during the 1976 and 1982-1983 El Niño events. Coles and Reinecker (2001) suggested that interior exchange had a seasonal cycle associated with seasonal variability of pycnocline thickness due to the seasonality of local Ekman pumping. Huang and Wang (2002) examined interannual variability of interior transport and found an ENSO-related signal. McPhaden and Zhang (2002) analyzed hydrographic data in the past 50 years and found that pycnocline transport convergence into the tropical Pacific displays variations on multi-decadal time scales. From the 1970s to the 1990s, there has been less and less subtropical pycnocline water coming to the tropics. Together with the inference that the equatorial upwelling became weaker during this period due to weakening trade wind, they suggested that the STC had been slowing down since the 1970s. While the change of interior transport from the 1970s to the 1990s was significant, that through the western boundary was unclear (due to the relatively large uncertainty of LLWBC transport estimates).

Understanding the mean and variability of tropical-subtropical exchange can elucidate different underlying processes. One way through which tropical heat content is affected by the subtropics is through convergence (divergence) of oceanic meridional heat transport. Let

\[ VT(y) = \int_{xw}^{xe} \int_{-h}^{0} v(x, y, z) \cdot t(x, y, z) \, dx \, dz \]

denote temperature flux across a zonal section at latitude \( y \), where \( v \) and \( t \) are meridional velocity and temperature, \( x \) is longitude, and \( z \) is depth. The inner integral is along the vertical direction from a certain depth \(-h\) (bottom of the ocean or base of the pycnocline) to the surface. The outer integral is from the western to the eastern boundary (\( xw \) to \( xe \)). \( VT(y) \) represents meridional heat transport across latitude \( y \) when \( \int_{xw}^{xe} \int_{-h}^{0} v(x, y, z) \, dx \, dz = 0 \) (i.e., no net volume transport). The convergence or divergence of \( VT \) between two latitudes \( y1 \) and \( y2 \), \( \Delta VT = VT(y2) - VT(y1) \), results in a change of heat content between these two latitudes and above \(-h\). If one decomposes the meridional flow and temperature into time mean and its anomaly:

\[ v(x, y, z) = \bar{v}(x, y, z) + v'(x, y, z) \quad \text{and} \quad t(x, y, z) = \bar{t}(x, y, z) + t'(x, y, z), \]
the heat transport can be expressed as

$$VT(y) = \int_{z_1}^{z_2} \int_{x_1}^{x_2} (\bar{v} + \bar{v}' + v' \bar{i} t + v' t') dx dz$$

where longitude, latitude, and depth indicators inside the integrals have been omitted for simplicity. The first term, the advection of mean temperature by mean flow, results in a mean heat transport. The second and third terms, corresponding to advection of temperature anomaly by mean flow and of mean temperature by anomalous flow, could result in heat transport anomaly depending on the time scale of $v'$ or $t'$. The last term, the advection of temperature anomaly by anomalous flow, may contribute to the mean and/or the variability of heat transport depending on the correlation of $v'$ and $t'$ and its variability in time.

Decadal temperature anomaly $t'$ has been observed to propagate from the subtropics towards the tropics from the late 1970s to the late 1980s (Deser et al. 1996, Zhang et al. 1999). Gu and Philander (1997) hypothesized that the advection of decadal $t'$ originating from the subtropics by $\bar{v}$ could affect sea surface temperature (SST) in the equatorial Pacific and hence tropical wind to cause decadal oscillation or decadal modulation of ENSO. However, decadal $t'$ cannot be traced as a coherent pattern all the way to the equatorial Pacific either in observations or in models (examples of the latter include Schneider et al. 1999 and Hazeleger et al. 2001). This casts an uncertainty about the role of $\bar{v} t'$ in affecting upper-ocean temperature of the equatorial Pacific.

Meanwhile, Kleeman et al. (1999) proposed an alternative hypothesis of heat transport mechanism: $\bar{i}$ advected by $v'$ (reflecting the oscillation of the STC on decadal time scales). Recent analysis by McPhaden and Zhang (2002) has provided observational evidence of decadal $v'$, with the strength of the STC decreasing from the 1970s to the 1990s, consistent with the tendency of tropical heat content during this period. There has not been any theoretical or observational work related to $v' \bar{i} t'$ on decadal time scales, in part due to the lack of observational evidence. However, this term may be significant on interannual time scales.

Understanding $v'$ is important to the assessment of the potential role of $v' \bar{i} t'$ Since meridional heat transport involves a zonal integration of this product, it is necessary to investigate the zonal structure of $v'$, especially in terms of the relative variation near the western boundary and in the interior. This is the focus of the present study. Specific questions to be addressed are: (1) how
does \( v' \) vary on interannual and decadal time scales in the boundary and interior of the pycnocline? (2) how are they different from the time-mean picture? (3) what are the forcing mechanisms responsible for the variability? To address these issues, we analyze the simulation by a state-of-the-art global ocean general circulation model (OGCM) for the period of 1980 to 2000 along with satellite-derived sea level and wind data and a data assimilation product in the 1990s. Variations of pycnocline transport via the boundary and interior pathways are quantified. Their relations to wind forcing and to the Indonesian throughflow (ITF) are discussed. The configuration of the model and its mean state are described in the next section. Interannual and decadal variations of boundary and interior pycnocline transports are presented in section 3. Forcing mechanisms are investigated in section 4 where relations of pycnocline transport to off-equatorial wind stress curl, equatorial zonal wind stress, and the Indonesian throughflow are examined; several sensitivity experiments are performed to help isolate the effect of individual forcings. The findings are summarized in section 5.

2. Model configuration and mean state

The OGCM of the Massachusetts Institute of Technology (Marshall et al. 1997a and 1997b) is used for this investigation. The model configuration is the same as that adopted by Lee et al. (2002) except that the background mixing coefficients are somewhat smaller here. The model domain is global in zonal direction and spans from 75°S to 75°N meridionally. The horizontal resolution is 1°×1° poleward of 23°N(S), telescoped to 1°×0.3° in the tropics. There are 46 vertical levels with a thickness of 10 m in the upper 150 m, gradually increasing to 400 m at depth. Two advanced mixing schemes are used by the model: the so-called KPP vertical mixing (Large et al. 1994) and the GM-Redi isopycnal mixing (Redi 1982, Gent and McWilliams 1990). The model is forced at the surface by 12-hourly wind stress and daily heat and freshwater fluxes. These fluxes are based on the reanalysis product of the National Center for Environmental Prediction (NCEP) (Kalnay et al. 1996) except that the time averages are replaced by a COADS product (da Silva et al. 1994). In addition to the imposed surface heat and freshwater fluxes, model SST and sea surface salinity are relaxed towards NCEP's SST and Levitus'98 climatological mean salinity (Boyer and Levitus 1998), respectively, with time scales typically being 1-2 months.
After a 10-year spinup from rest with Levitus'98 climatological temperature and salinity (Boyer and Levitus 1998) forced by seasonal climatological forcings, the model is integrated using the forcings from 1980 to 2000. A more detailed description of the model configuration and comparison of the model state with satellite and in-situ data are provided by Lee et al. (2002) and so will not be repeated here. The remaining discussion in this section compares the mean pycnocline structure and transports of the model with observational based analyses, focusing primarily near 10°N and 10°S which are important to tropical-subtropical exchanges.

Figure 1 shows the zonal sections of averaged potential density from the model and from Levitus98 data at 10°N and 10°S. The large-scale representation of the density structure is reasonable both in the western boundary and interior. The model’s pycnocline is somewhat sharper than that of the data in the eastern part of the basin. The averaged depths of middle-to-lower pycnocline (e.g., $\sigma_\theta=24-26.5$ kg/m$^3$) in the model are shallower than those in the data by about 10 m at 10°N and 20-30 m at 10°S. In the present study, pycnocline transport is defined as the volume transport between 22-26.5 potential-density ($\sigma_\theta$) surfaces. When the $\sigma_\theta=22$ surface outcrops, we treat the depth of 50 m as the top of the pycnocline so as to exclude Ekman transport approximately (but nevertheless excluding some near-surface geostrophic flow as well).

To illustrate the spatial distribution of boundary and interior flow in the pycnocline, we define a “pseudo” stream function for pycnocline transport $\tilde{\psi}(x, y) = \int_x^e V(x, y)dx$ (i.e., total meridional transport in the pycnocline $V(x, y)$ at latitude $y$ integrated from the eastern boundary $xe$ to longitude $x$). It is not a stream function in the non-divergent sense because of non-zero net pycnocline transport across a latitude. The time-mean $\tilde{\psi}$ (Figure 2) illustrates pycnocline flow through the interior and along the western boundary associated with the LLWBCs, namely, the MC in the north and the NGCU in the south. The main features are consistent with previous studies: (1) the separation of boundary and interior flow is more distinguishable in the northern than the southern hemisphere; (2) there is more interior flow into the tropics from the southern subtropics than from the northern one; (3) flow through the western boundary in the northern hemisphere is significantly larger than that through the interior; and (4) interior flow in the northern hemisphere is
more confined to the central part of the basin. The weaker interior flow in the northern hemisphere is related to the so-called “potential vorticity barrier” associated with the region of high potential vorticity under the Intertropical Convergence Zone (ITCZ) which limits the averaged amount of interior flow reaching the tropics from the northern subtropical gyre on average. The circulation patterns illustrated by $\tilde{\psi}$ is qualitatively similar to the geostrophic streamlines presented by McPhaden and Zhang (2002) (their Figure 1a) for the center of the pycnocline.

The western-boundary and interior transports at 10°N are defined as pycnocline transports from the western boundary to 130°E and from this longitude to the eastern boundary, respectively. The separation of western boundary and interior at 10°S is chosen to be 159°E. The model’s mean boundary and interior pycnocline transports are 15.7 and 5.6 Sv (1 Sv = $10^6$ m$^3$/s) for 10°N and 9.2 and 10.1 Sv for 10°S. The boundary transport at 10°N is close to but somewhat larger than various estimates based on analyses of in-situ data: 13 Sv by Lukas et al. (1991) based on ADCP measurements, 15 Sv by Wijffles (1993) based on an inverse model of hydrography, 14 Sv by Huang and Liu (1999) based on the ocean analysis of temperature profile data using the NCEP ocean model. The interior transport at 10°N is close to the 5-Sv estimate by Johnson and McPhaden (1999) based on hydrography, and somewhat larger than the 4-Sv value obtained by Wijffles (1993) and Huang and Liu (1999).

Transports at 10°S are several Sv smaller than the observational-based analyses (Lindstrom et al. 1987, Butt and Lindstrom 1994, Johnson and McPhaden 1999, and Huang and Liu 1999), especially that through the western boundary. The choice of the mean wind product is believed to affect pycnocline transport. Background mixings may also matter because they affect horizontal density gradient and thus vertical distribution of volume transport. In addition, the realism of island geometry off the coasts of New Guinea and New Ireland in our model is questionable because of the limited resolution. Lindstrom et al. (1987) and Butt and Lindstrom (1994) identified two LLWBCs in the South Pacific, the New Guinea and New Ireland Coastal Undercurrents, carrying about 7 and 5 Sv of transport, respectively. In our solution, the distinction of these two currents is not clear because various straits in the region are not adequately represented. However, the boundary current transports reported by Lindstrom et al. (1987) and
Butt and Lindstrom (1994) are based on fairly short temporal records. It is not clear whether the reported transports are representative of long-term averages. While we are working towards improving the model's mean state, it is the variability of pycnocline transport that is the focus of the present study. As will be discussed, the variability of pycnocline transport is primarily controlled by the variability of wind forcing. Therefore, the findings with regard to variability may not be strongly subject to the limitation of mean transports at 10°S.

3. Interannual-to-decadal variations of pycnocline transport

Anomalies of boundary, interior, and net pycnocline transports (i.e., with time mean removed) are shown in Figure 3. Two features are seen: the variation of boundary transport tends to be anti-correlated to and generally has a smaller magnitude than that of interior transport. The correlation coefficients are -0.73 for 10°N and -0.93 for 10°S, with both being significant at 95% confidence level. The standard deviation of pycnocline transports through the boundary and interior as well as their sum are listed in Table 1. The compensation of interior flow by boundary flow at 10°S is larger than that at 10°N. Let $\sigma_i$ and $\sigma_n$ represent standard deviations of interior and net pycnocline transports, respectively. Then the ratio $C=(\sigma_i-\sigma_n)/\sigma_i$ can be used to characterize the degree of compensation of interior flow by boundary flow. The value of $C$ is 31% at 10°N and 62% at 10°S. The partial compensation of interior by boundary flow not only occurs on interannual but on decadal time scales as well. Figure 4 shows anomalies of boundary and interior transport after a 5-year smoothing has been applied. There is less subtropical pycnocline water intrusion into the tropics in the 1990s than in the 1980s. The opposite is true for boundary transport anomaly.

Based on analysis of hydrography data during the past 50 years, McPhaden and Zhang (2002) found that equatorward convergence of interior pycnocline flow into the 9°N-9°S band had been decreasing from the 1970s to the 1990s. A similar decrease was found in the convergence into the 15°N-15°S band. Although we do not have the simulation for the 1970s, the mean difference of interior transport between the 1990s and 1980s in our model, 2.9 Sv, is comparable to the estimate by McPhaden and Zhang (2002) of 3-4 Sv (with an uncertainty of 1-2 Sv). In their analysis, the estimated boundary transport was fairly uncertain (several Sv) due to insufficient data. In the
present study, decadal difference in boundary transport at 10°N is –1.5 Sv (more subtropical pycnocline water intrusion into the tropics in the 1990s). So approximately 50% of the decadal change of interior flow is compensated by an opposite change of boundary flow at this latitude. As will be seen in the next section (in conjunction with Figure 10d), compensation on decadal time scales is found for the latitude band of approximately 9°-15°N. This finding prompts for additional measurements in the boundary to resolve decadal variability. On average, much of the Indonesian throughflow (ITF) is fed by the MC. Does the 1.5-Sv decadal change in pycnocline transport associated with the MC affect the Indian Ocean through the ITF or does it influence in the tropical Pacific only? In our model, decadal change in ITF transport is only 0.1 Sv, suggesting that the effect of decadal variation in pycnocline flow at 10°N is on the tropical Pacific.

At 10°S, a near decadal signal is also seen in the interior pycnocline transport (Figure 4b). Also evident is the compensation by boundary flow. However, the phase of the decadal signal of interior or boundary transport at 10°S leads the one at 10°N by about 3 years. The difference of interior pycnocline transport at 10°S between the 1990s and 1980s is 0.4 Sv. This is smaller than the corresponding estimate of about 3 Sv by McPhaden and Zhang (2002). Note that the error bars for their estimate in the southern hemisphere are larger than those in the northern hemisphere. Therefore, the difference from their estimate is not all that significant. Furthermore, as McPhaden and Zhang noted, there are differences in decadal variability among different wind products. It is not clear whether a different wind product would result in a larger decadal change or not.

Figure 5 shows zonal sections of meridional velocity of the LLWBCs at 10°N and 10°S averaged annually from 1997 to 2000. The MC is strongest in 1997, an El Niño year, and weaker from 1998-2000, which is more characterized by La Niña condition. The core depth of the MC is shallower in 1997 than it is in 1998-2000. The NGCU also displays similar interannual changes as that of the MC. However, the phase lags that of the MC by about a year (strongest and shallowest in 1998 instead of in 1997). As will be discussed in section 4, this is related to the phase difference of wind stress curl anomalies near 10°N and 10°S. The generally shallower core depth of the LLWBCs during predominant El Niño condition reflects the large-scale zonal tilt of the pycnocline as the westerly wind anomaly during El Niño raises and depresses the pycnocline in the western
and eastern part of the basin, respectively. Vice versa for La Niña. The cause for the variation in the strength of the MC, as discussed in the next section, is off-equatorial wind stress curl.

As a way to assess the accuracy of the model simulation, interannual changes of the LLWBCs are compared with those inferred from a data assimilation product in which sea level anomalies obtained by the TOPEX/Poseidon altimeter have been assimilated. The assimilation product, generated at the Jet Propulsion Laboratory (JPL) by the adjoint method using the same model, is available through a live-access data server http://eyre.jpl.nasa.gov/las. It is part of the collaborative effort by the Consortium “Estimating the Circulation and Climate of the Ocean” (ECCO; http://ecco-group.org). An introduction of the consortium activity is given by Stammer et al. (2002). Figure 6 shows the estimated meridional velocity sections at 10°N(S) for the LLWBCs from 1997 to 2000 (the counterpart of Figure 5, but from the assimilation). The variability in the simulation is qualitatively similar to that in the assimilation. However, changes in the strength and core depth of the LLWBCs from El Niño to La Niña in the simulation are not as large as those inferred from the assimilation. This is because the simulation underestimates the interannual change of tropical zonal wind and consequently the variation of the zonal tilt of the pycnocline. In fact, the magnitude of sea level anomaly associated with the 1997 El Niño in the simulation is too small because the westerly wind anomaly in 1997 (from the NCEP product) is too weak. The assimilation of sea level data results in more realistic anomalies of westerly wind and sea level (and thus more reliable variability in the core depth and circulation of the pycnocline).

Decadal variation of the MC is shown in Figure 7. It is stronger in the 1990s than it is in the 1980s. Its core depth is shallower in the 1990s when it is stronger. These features are somewhat similar to the difference in the characteristics of the MC El Niño and La Niña years as shown in Figures 5 and 6. The easterly trade in the equatorial Pacific is known to be weaker in the 1990s than the 1980s. The pycnocline is expected to become shallower in the west and deeper in the east in the 1990s in response to the change of tropical wind. Therefore, the shallower core depth of the MC in the 1990s reflects large-scale tropical zonal wind in a way that is similar to interannual changes associated with El Niño and La Niña. The decadal change in the core depth of the MC is consistent with the decadal change in the tilt of pycnocline depth reported by McPhaden and Zhang (2002).
There are insufficient in-situ data to characterize interannual variation of pycnocline flow, especially those via the LLWBCs. However, sea level observations such as those obtained by the TOPEX/Poseidon altimeter allow one to infer the variability of near-surface meridional geostrophic flow which is relatively coherent with the flow in the deeper part of the pycnocline (as seen from Figures 5 and 6). In fact, Lukas (1988) used in-situ sea level data to examine the variability of the MC. He found that interannual change in the strength of MC was anti-correlated with the equatorial heat content during the 1976 and 1982 El Niño events.

Figure 8 shows time series of anomalous east-west difference in sea level across the western boundary and across interior for the model, TOPEX-Poseidon data, and the assimilation. The respective time averages for the period of 1997-2000 have been removed to facilitate the comparison with the assimilation product available for this period. They are proportional to the magnitude of near-surface geostrophic transport (with reversed sign in the southern hemisphere). Both the model and the data suggest that the variability of near-surface geostrophic flow in the boundary is (1) anti-correlated to and (2) generally smaller than that in the interior. The magnitude of variability in the model simulation is smaller than that of the data at 10°N. At 10°S, the model also underestimates the variability since 1997, but overestimates that during 1994-1996. These differences are believed to be related to the fidelity of the wind stress product (NCEP) in representing actual wind variability. The differences between the assimilation and the data are much smaller. The similarity between the model simulation and the data indicates that the model has a reasonable skill in simulating the relative variability of pycnocline flow in the boundary and interior.

The above findings highlight two major differences between the mean and variability of tropical-subtropical mass exchange in terms of boundary and interior pathways. First of all, time-mean boundary and interior transports \( \overline{V}_b \) and \( \overline{V}_i \) are both equatorward and thus re-enforce the net exchange. However, the variability of boundary transport \( V'_b \) is generally out of phase with interior transport \( V'_i \) and thus mediates the net exchange. Secondly, \( \overline{V}_b \) is substantially larger than \( \overline{V}_i \) at 10°N whereas \( V'_b \) is generally smaller than \( V'_i \). In other words, boundary transport is more important to time-mean mass exchange while interior transport plays a more significant role in the variability of mass exchange.
4. Forcing mechanisms

Forcing mechanisms responsible for the variability of boundary and interior transports are examined in this section. In particular, we investigate processes responsible for the counteracting tendency of boundary and interior pycnocline transports and the larger variability of the latter. The importance of three possible forcings is evaluated: (1) equatorial zonal wind stress, (2) off-equatorial wind stress curl, and (3) Indonesian throughflow.

Using a reduced-gravity model of the tropical Pacific forced by observed wind for the period of 1970-87, Zebiak (1989) found that meridional transports through the western boundary at 5°N and 5°S generally opposed the tendency of equatorial heat content, but were more than compensated by the interior transports. They suggested that the tendency was consistent with the reflection of equatorial long Rossby waves (Cane and Sarachik 1983). Equatorial westerly wind generated eastward-propagating, downwelling Kelvin waves. Associated with these waves was an eastward equatorial jet. Water carried by this jet returned poleward further east in the interior. Meridional mode of westward-propagating, upwelling Rossby waves was also excited by the equatorial westerly wind. When these waves impinged on the western boundary, they produced convergence of flow towards the equator, which compensated for waters carried eastward by the equatorial jet. Anti-correlated boundary and interior transports at 5°N(S) were also found in a similar model by Springer et al. (1990) for the period of 1979-1983. They suggested that the convergence of boundary flow appeared to be related to the reflection of equatorial Rossby waves only after the onset of the 1982-83 El Niño event, but not before. However, transports at 10°N and 10°S were not examined in these studies.

Consistent with the simulations by these reduced-gravity models, counteracting tendencies between boundary and interior flow are seen at 5°N(S) in our model. Figure 9 compares boundary, interior, and net pycnocline transports at 5°N with the corresponding ones at 10°N. Net pycnocline flow across 5° and 10°N (panel c), being the lower branch of the STC at two different latitudes, are close to each other. It suggests that diapycnal convergence/divergence of pycnocline water and the
change in the volume of pycnocline water between the latitudes due to along-isopycnal flow are small on interannual and decadal time scales. Interior transports between these two latitudes are also somewhat coherent (panel b). However, boundary transports across the two latitudes are not similar at all (panel a). On interannual time scales, there is little coherence at any time lag between boundary transports at 5° and 10°N. On decadal time scales, the tendencies of the two are actually opposite. If boundary transports at 5° and 10°N are both associated with the reflection of Rossby waves excited by equatorial zonal wind, one would expect the variability at 10°N to be substantially smaller than that at 5°N because the amplitude of equatorial Rossby waves decreases substantially from 5° to 10°N. In the model, however, the magnitudes of boundary transport variability at these two latitudes are comparable. These features suggest that the reflection of equatorial Rossby waves, which presumably affects boundary transports at 5°N significantly, may not be very important at 10°N.

To investigate changes in pycnocline circulation that affect transports at 5°N(S) and 10°N(S), interannual and decadal changes of “pseudo” stream function $\tilde{\psi}(x, y)$ (defined in section 2) are shown in Figure 10. Two patterns of anomalous circulation affect transports at 10°N(S): one that originates in the equatorial band and one that centers at latitudes near and poleward of 10°N(S). The former is more easily seen in the change from 1996 to 1997 (panel a), characterized by intense circulation anomaly in the equatorial band with maxima near 3°N and 4°S and with the intensity decreasing further away from the equator. This is reminiscent of the effect of equatorial Rossby waves, which have maximum amplitude near these latitudes. On top of this pattern is an anomalous circulation centered between 10° and 15°N west of the dateline. This off-equatorial pattern changes direction from 1997 to 1998 (panel b) and from 1998 to 1999 (panel c). Off-equatorial anomaly of circulation is also found in the southern hemisphere (more apparent in panels b and c). While that near 10°N is more zonally oriented, the one near 10°S is tilted in the northwest-southeast direction. These off-equatorial circulation anomalies have a larger influence on transports at 10°N(S) than the patterns centered in the equatorial band. Similar features of anomalous pycnocline circulation are found during the 1981-1984 period (not shown) that features the development of another large El Niño and the subsequent La Niña events.
Change in pycnocline circulation from the 1980s to the 1990s (Figure 10d) is characterized by an off-equatorial pattern centered between 10° and 15°N in the western to central Pacific. It resembles the difference between 1997 and 1998 (Figure 11b, but with opposite direction). Figure 10d also explains why the decadal tendencies of boundary transport at 5° and 10°N are different. The anomalous interior flow across 10°N in the 1980s relative to the 1990s does not reach the equatorial band through the interior, but connects with anomalous boundary flow at 5°N.

The pycnocline circulation patterns indicate that, both on interannual and decadal time scales, change in off-equatorial circulation affects boundary transports at 10°N(S) significantly. The direction of off-equatorial anomalous circulation near the western boundary in the pycnocline is somewhat similar to that of full-depth integrated circulation in many cases (Figure 11). This consistency suggests that, to some extent, the change in pycnocline circulation near the western boundary reflects the change of full-depth integrated horizontal circulation. As discussed in the following, the change in horizontal circulation is found to be primarily a result of Ekman pumping by off-equatorial wind stress curl in the western part of the basin.

Time-mean wind stress curl and its zonal average are presented in Figures 12a and b. Their local maxima and minima occur near 10°N and 10°S, respectively, both resulting in local maxima in Ekman pumping (defined as wind stress curl divided by planetary vorticity that is negative in the southern hemisphere). These latitudes roughly correspond to the bifurcation points of the NEC and SEC. While the curl pattern in the northern hemisphere is more zonally oriented, that in the southern hemisphere is tilted along the northwest-southeast direction. Interannual-to-decadal variability of the curl (Figures 12c and d) has local maxima near 15°N and 10°S (panels c and d), primarily due to curl anomalies over the western part of the basin. These anomalies are caused by meridional movement and the change in the strength of the curl pattern. An example is shown in Figure 13 for the period of 1997 to 1999 both for the NCEP product (panels a, c, and e) and for ERS-2 scatterometer measurement (b, d, and f). The band of maximum wind stress curl near the western Pacific near 10°N is tilted upward in 1997, downward in 1998, and upward again in 1999 (as marked by the black line segments). Changes in the pattern of wind stress curl resemble those of the ITCZ and the South Pacific Convergence Zone (SPCZ). A similar feature is seen on decadal time scales for the NCEP product (not shown).
To illustrate the variation of off-equatorial wind stress curl and the corresponding change in oceanic state, anomalies of wind stress curl and sea level for the period of 1997-1999 (referenced to the mean of this period) are shown in Figure 14. In 1997, the negative sea level anomaly near 10°-15°N and west of 160°E corresponds to the positive curl anomaly primarily west of the dateline. The positive sea level anomaly west of 160°E developing from 1998 to 1999 corresponds to the negative curl anomaly during the same period. In the southern hemisphere, sea level anomaly is related to wind stress curl anomaly of the same sign because planetary vorticity is negative south of the equator. Near 10°S, sea level anomaly changes from being negative in 1998 to being positive in 1999, corresponding to predominately negative (positive) curl anomaly in 1998 (1999). The anomalies near 10°N tend to be more zonal while those near 10°S are tilted along the northwest-southeast direction. These characteristics are consistent with the orientations of the anomalous patterns of off-equatorial circulation shown in Figure 10. The co-variability of wind stress curl and sea level in the model is generally consistent with wind measurement by the ERS-2 scatterometer and sea level measurement by the TOPEX/Poseidon altimeter (Figure 15). Nevertheless, the NCEP product underestimates the magnitude of wind stress curl anomaly near 10°N(S), giving rise to smaller variation of sea level.

In principle, the variability of sea level in the western Pacific reflects the integrated effect of wind stress curl across the basin. However, the anomaly of wind stress curl over the western part of the basin near 10°N(S) is much larger than that in the east such that the variability of sea level and pycnocline flow in the western Pacific are largely forced locally. In fact, boundary pycnocline transports at 10°N(S) are reasonably well correlated to “local” Sverdrup transports (with sign reversed) computed from curl anomalies over the western part of the basin (Figure 16), suggesting the important role of these wind stress curl anomalies in causing the variability of boundary flow. The magnitude of pycnocline transports are generally smaller than the “local” Sverdrup transports because the former is only part of the full-depth integrated transport inferred from the Sverdrup relation.

The coherent changes in pycnocline circulation, sea level, and wind stress curl indicate that local wind stress curl anomalies near 10°N(S) modify the strength of the horizontal circulation to create
counteracting tendency of boundary and interior transport in the pycnocline. This effect is further examined through a sensitivity experiment. The control experiment is the last year of the spinup with seasonal forcing. A wind perturbation is introduced in the perturbation experiment. The zonal and meridional wind perturbation have the form of $\tau_x = \partial P/\partial y$ and $\tau_y = \partial P/\partial x$, where $P$ is a Gaussian “wind stress potential” with an e-folding scale of $5^\circ$ and $100^\circ$ in meridional and zonal directions, respectively. The pattern is centered at $190^\circ$E, $10^\circ$N. The effective magnitude of the wind stress curl perturbation is $2 \times 10^{-8}$ N/m$^3$, close to the observed variability of wind stress curl near $10^\circ$N. The perturbed boundary and interior transports integrated over the entire water column and in the pycnocline are coherent (Figure 17), both showing a counteracting tendency that reflects the change in the strength of horizontal circulation. This is consistent with the fact that the directions of anomalous pycnocline and barotropic circulation near the western boundary generally agree with each other as discussed earlier in conjunction with Figures 10 and 11. The perturbed pycnocline transports through the boundary and interior more or less cancel out.

In the 1980-2000 model simulation, changes in boundary and interior pycnocline transports do not cancel out. There is a net pycnocline transport because interior flow is larger than that through the boundary. This suggests a second forcing mechanism of the pycnocline flow, in particular, one that affects the net pycnocline transport. On longer time scales, net pycnocline transport (the lower branch of the STC) should approximately balance the near-surface Ekman flow (the upper branch of the STC). Near-equatorial zonal wind is known to vary on interannual and decadal time scales associated with ENSO and decadal variability. This would affect the strength of the STC and thus net pycnocline transport. Net pycnocline transport at $10^\circ$N(S) is compared with Ekman transports computed from zonal wind stress averaged between the equator and $10^\circ$N(S) (Figure 18). They are reasonably well correlated even on interannual time scales, reflecting the response of the lower branch of the STC to near equatorial zonal wind on these time scales. Generally speaking, there is less equatorward net pycnocline transport during El Niño when the trade wind is more westerly, corresponding to weaker poleward Ekman flow. The opposite is true during La Niña.

To further examine the effect of near-equatorial zonal wind on net pycnocline transport, another perturbation experiment is performed in which the wind field is perturbed by a time-constant, globally uniform westerly wind. The magnitude of the westerly is close to the interannual-to-
decadal standard deviation of zonal wind stress averaged over the tropics. The choice of the uniform westerly wind isolates the effect of zonal wind without introducing off-equatorial curl. Both at 10°N and 10°S, perturbed interior transport is found to be larger than that of the boundary (Figure 19), giving rise to a net pycnocline transport. There is small boundary transport in opposite direction to the interior flow. It indicates that equatorial zonal wind anomaly without any change in off-equatorial curl would also have some influence, although small, at latitudes as far as 10° away from the equator. This is not inconsistent with the convergence of boundary flow associated with Rossby waves impinging at the western boundary discussed at the beginning of this section.

The variability of off-equatorial wind stress curl near 10°N(S) is correlated to near-equatorial zonal wind stress (Figure 20). The general tendency is that when near equatorial zonal wind is more westerly, wind stress curl near 10°N(S) tends to be more positive (negative), both creating anomalous low sea level near the western boundary and anomalous equatorward pycnocline flow in the boundary. Figure 21 illustrates the combined effect of the two forcings (northern-hemisphere example) in creating counteracting boundary and interior pycnocline transports, with the latter being dominant. In summary, off-equatorial wind stress curl modifies the strength of horizontal circulation to create counteracting boundary and interior transports in the pycnocline that have comparable magnitudes; near-equatorial zonal wind affects the strength of the STC and thus the net pycnocline transport (mostly distributed in the interior). The combined effect is a larger interior transport than boundary transport.

The proposed forcing mechanism can explain not only interannual variability, but decadal variability of boundary and interior pycnocline flow at 10°N as well. This is because equatorial zonal wind stress and off-equatorial wind stress curl co-vary both on interannual and decadal time scales. On interannual time scales, when equatorial zonal wind stress is more westerly (e.g., associated with El Niño), the ITCZ tends to shift equatorward while its intensity changes. These changes result in a curl anomaly near 10°N. A similar behavior is seen on decadal time scales. Equatorial zonal wind stress is more westerly in the 1990s than it is in the 1980s (solid curve in Figure 22a). Associated with this is a more positive difference of wind stress curl between 10° and 15°N (Figure 22b) and a cyclonic anomaly of pycnocline circulation in the western Pacific centering near 12°N (Figure 10d). These features resemble the difference between El Niño and La Niña conditions.
Niña years. For instance, near-equatorial zonal wind in 1997 is more westerly than that in 1998 west of 140°W (dashed curve in Figure 22a); this corresponds to a more positive curl in 1997 west of the dateline (Figure 22c) and a more cyclonic pycnocline circulation in the western Pacific centered near 12°N in 1997 than in 1998 (i.e., the pattern seen in Figure 10b but with reversed direction, because the difference between 1998 and 1997 is shown in Figure 10b). The anomaly of westerly wind in the 1990s (relative to the 1980s) spans from 150°E to 110°W while that associated with the 1997 El Niño is west of 140°W (Figure 22a). A similar difference in zonal extent is seen for the anomalies of wind stress curl between interannual and decadal time scales (Figures 22b and c). The anomalous patterns of pycnocline circulation on interannual and decadal time scales (Figure 10) also display such a difference in zonal extent.

According to the proposed forcing mechanism, the relative contributions of off-equatorial wind stress curl and near-equatorial zonal wind stress would dictate the relative variability of boundary and interior transports. The larger the off-equatorial wind stress curl, the stronger the pycnocline recirculation and thus the compensation between boundary and interior flow. On the other hand, the larger the near-equatorial zonal wind stress, the more dominant the variability of interior flow becomes. At 10°N, the variability of local Sverdrup transport associated with off-equatorial curl is comparable to that of Ekman transport inferred from near-equatorial zonal wind (Figure 20a), with the corresponding standard deviations being 2.9 and 2.5 Sv, respectively. At 10°S, however, the former is more dominant than the latter (Figure 20b) because the corresponding standard deviations become 4.8 and 2.0. It indicates that local wind stress curl is a more dominant forcing for 10°S than for 10°N. This is consistent with the finding that the compensation of boundary and interior pycnocline transports at 10°S is larger than that at 10°N.

The above discussion focuses on wind forcings. The possible role of the ITF is discussed in the following. Using the same OGCM with somewhat different background mixing coefficients, Lee et al. (2002) found that the time-mean strength of the NGCU is reduced significantly when the ITF is blocked off. This is because the circulation loop that goes around Australia has been cut off. The effect on time-mean strength of the MC, however, is relatively small because the MC is north of this circulation loop. Does the ITF affect the variability of tropical-subtropical exchange?
To address this issue, we analyze boundary and interior transports obtained from the two simulations performed by Lee et al. (2002) with and without the ITF (Figure 23). The counteracting tendency between boundary and interior flow and the larger magnitude of the latter are found to be insensitive to the presence/absence of the ITF. Even the decadal signal at 10°N is unaffected by the ITF. At 10°S, the net pycnocline transport drifts away when the ITF is blocked off. The open-closed ITF experiments are somewhat idealized because the wind field is assumed to be unaffected by the presence (or absence) of the ITF. In reality, longer-term variation in the ITF may change air-sea heat flux, especially in the southern Indian Ocean (Hirst and Godfrey 1993, Lee et al. 2002), which may modify the wind over the Pacific Ocean through atmospheric teleconnection to affect tropical-subtropical exchange in that ocean. This is beyond the scope of the present study because it requires sensitivity experiments using a coupled ocean-atmospheric model.

5. Summary and Conclusions

Interannual-to-decadal variations of tropical-subtropical mass exchange in the Pacific Ocean is investigated using a near-global ocean general circulation model for the period of 1980 to 2000 along with sea level and wind measurements by satellite altimeter and scatterometer and a data assimilation product in the 1990s. The analysis focuses on the variability of pycnocline transports through the western boundary and interior across 10°N and 10°S, the approximate latitudes at which the NEC and SEC bifurcate into mid- and low-latitude western boundary currents.

In contrast to time-mean exchange where boundary and interior pycnocline transports are both equatorward, the variations of boundary and interior pycnocline transports are found to be generally anti-correlated to each other. Moreover, the variation of boundary pycnocline transport is smaller than that of the interior. This is again different from the time-mean exchange where the boundary transport at 10°N is substantially larger than that through the interior. Interannual variations of boundary and pycnocline transports are consistent with near-surface geostrophic flow inferred from sea level data collected by the TOPEX/Poseidon altimeter. Equatorward interior pycnocline flow is weaker in the 1990s than it is in the 1980s, consistent with recent analysis of observational data.
Nevertheless, approximately half of it is compensated by an opposite change in the boundary flow. The result indicates that interior exchange in the North Pacific is more important to interannual and decadal variability in the tropical Pacific despite the fact that the western boundary current (the Mindanao Current) carries more subtropical pycnocline water to the tropics than the interior does on the time mean.

The counteracting tendency of the boundary and interior flow and the more dominant variation of the latter are attributed to the combined effect of variations associated with off-equatorial wind stress curl primarily over the western Pacific and near-equatorial zonal wind stress. These two forcings are correlated on interannual and decadal time scales. The former changes the strength of the horizontal circulation and results in variation of boundary pycnocline flow that is opposite in direction and comparable in magnitude to that of the interior pycnocline flow. The latter primarily affects the strength of the shallow meridional overturning circulation with primarily interior pycnocline flow opposing the surface Ekman flow. The proposed forcing mechanisms help explain the differences in variability between 10°N and 10°S. The larger compensation of boundary and interior flow at 10°S (than 10°N) is consistent with the larger variability of wind stress curl near 10°S. Namely, the variation in the strength of horizontal circulation dominates. The more dominant variability of interior than boundary transport at 10°N is consistent with the larger variation of zonal wind stress in the northern tropical Pacific. In other words, the variation in the strength of the meridional circulation is more important. Although the presence/absence of the Indonesian throughflow affects time-mean boundary transport at 10°S significantly, it does not affect the relative variability of boundary and interior pycnocline transports.

Various aspects related to the variability of boundary and interior pycnocline flow of the model are compared with satellite data, inferences from in-situ observation, and data assimilation. The accuracy of the variability of pycnocline flow simulated by the model depends much on how realistic the wind product is. Smaller variability of the NCEP wind in the late 1990s compared to satellite scatterometer measurement results in generally weaker variability of boundary and interior flow simulated by the model in comparison with satellite altimeter data. To the first order, however, the model has fidelity in describing the relative variability of boundary and interior pycnocline flow.
Acknowledgement

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References:


Table 1. Interannual-to-decadal standard deviation of pycnocline transport through the western boundary and interior

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Figure captions

Figure 1. Zonal sections of time-mean potential density: (a) model at 10°N, (b) Levitus data 10°N, (c) model at 10°S, and (d) Levitus data at 10°S. Contour unit is kg/m³.

Figure 2. Time-mean pseudo stream function of pycnocline transport (see text for definition). Contour unit is Sv.

Figure 3. Anomalies of pycnocline transports at (a) 10°N and (b) 10°S.

Figure 4. Same as figure 3 except that a five-year smoothing has been applied.

Figure 5. Zonal sections of annually averaged meridional velocity of the model at (a-d) 10°N and (e-h) 10°S from 1997 to 2000. Contour unit is cm/s.

Figure 6. Same as figure 5 except that they are from a data assimilation product.

Figure 7. Zonal sections of meridional velocity at 10°N averaged over (a) the 1980s, (b) the 1990s, and (c) their difference. Contour unit is cm/s.

Figure 8. Anomalies of east-west sea level differences across the western boundary and across the interior at (a) 10°N and (b) 10°S for model simulation, assimilation, and TOPEX/Poseidon data.

Figure 9. Comparison of (a) boundary, (b) interior, and (c) net pycnocline transports at 10°N with those at 5°N.

Figure 10. Interannual and decadal changes of pycnocline transport pseudo stream function: (a) from 1996 to 1997, (b) from 1997 to 1998, (c) from 1998 to 1999, and (d) from the 1980s to the 1990s. Contour increment is 1 Sv.
Figure 11. Interannual and decadal changes of barotropic transport stream function: (a) from 1996 to 1997, (b) from 1997 to 1998, (c) from 1998 to 1999, and (d) from the 1980s to the 1990s. Contour increment is 2 Sv.

Figure 12. Time mean and interannual-to-decadal variability of wind stress curl (a and c) and their corresponding zonal averages (b and d). The unit is $10^{-7}$ N/m$^3$.

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Figure 16. Comparison of anomalous boundary pycnocline transport at (a) 10°N and (b) 10°S with local Sverdrup transport computed from wind stress curl averaged over the western Pacific.

Figure 17. Perturbed transports due to an imposed wind stress curl anomaly centered near 10°N. Perturbed boundary and interior transports are comparable.

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Figure 21. Schematic illustration of the combined effect of off-equatorial wind stress curl and near equatorial zonal wind stress in causing counteracting boundary and interior pycnocline transports and a larger magnitude of the latter (northern hemisphere example).

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Figure 23. Anomalies of pycnocline transports at (a and c) 10°N and (b and d) 10°S with the Indonesian channels (a and b) open and (c and d) closed.
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(c) Model 10°S

(d) Levitus 10°S

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