Journal of Oceanology and Limnology
Vol. 37 No. 2, P. 423-433, 2019
https://doi.org/10.1007/s00343-019-8027-y

The seasonal variation of the North Pacific Meridional Overturning Circulation heat transport*

LIU Hongwei[1, 2, 5], ZHANG Qilong[1, 5, **], PANG Chongguang[1, 5], DUAN Yongliang[3, 4], XU Jianping[2]

[1] *Key Laboratory of Ocean Circulation and Waves, Institute of Oceanology, Chinese Academy of Sciences and Qingdao National Laboratory for Marine Science and Technology, Qingdao 266071, China*
[2] *State Key Laboratory of Satellite Ocean Environment Dynamics, MNR, Hangzhou 310012, China*
[3] *Center for Ocean and Climate Research, First Institute of Oceanography, MNR, Qingdao 266061, China*
[4] *Laboratory for Regional Oceanography and Numerical Modeling, Qingdao National Laboratory for Marine Science and Technology, Qingdao 266061, China*
[5] *Center for Ocean Mega-Science, Chinese Academy of Sciences, Qingdao 266071, China*

Received Feb. 2, 2018; accepted in principle Apr. 4, 2018; accepted for publication Jun. 19, 2018

Abstract Based on the 50-year Simple Ocean Data Assimilation (SODA) reanalysis data, we investigated the basic characteristics and seasonal changes of the meridional heat transport carried by the North Pacific Meridional Overturning Circulation. And we also examined the dynamical and thermodynamic mechanisms responsible for these heat transport variability at the seasonal time scale. Among four cells, the tropical cell (TC) is strongest with a northward heat transport (NHT) of (1.75±0.30) PW (1 PW=1.0×10^{15} W) and a southward heat transport (SHT) of (-1.69±0.55) PW, the subtropical cell (STC) is second with a NHT of (0.71±0.65) PW and SHT of (-0.63±0.53) PW, the deep tropical cell (DTC) is third with a NHT of (0.18±0.03) PW and SHT of (-0.18±0.11) PW, while the subpolar cell (SPC) is weakest with a NHT of (0.09±0.05) PW and SHT of (-0.07±0.09) PW. These four cells all have different seasonal changes in their NHT and SHT. Of all, the TC has stronger change in its SHT than in its NHT, so do both the DTC and SPC, but the seasonal change in the STC SHT is weaker than that in its NHT. Therefore, their dynamical and thermodynamic mechanisms are different each other. The local zonal wind stress and net surface heat flux are mainly responsible for the seasonal changes in the TC and STC NHTs and SPC SHT, while the local thermocline circulations and sea temperature are primarily responsible for the seasonal changes of the TC, STC and DTC SHTs and SPC NHT.

Keyword: meridional overturning circulation; heat transport; North Pacific; seasonal variation

1 INTRODUCTION

The radiation, as a key component of the global climate system, is surplus in the tropics and deficient in the polar region. In order to maintain the heat balance of the whole earth system, the strong meridional heat transport must exist from the low latitude to high latitude (Trenberth and Solomon, 1994). And ocean circulation plays a vital role in this process. Namely, the ocean circulation conveys low-latitude warm water to the high latitude, and meanwhile releases a lot of heat to the atmosphere overlying them, and then the atmospheric circulations also transfer the heat to the higher latitude (Hsiung, 1985). In the North Pacific, the poleward heat transport is mainly completed by the western boundary current (such as the Kuroshio) and the Ekman flow in the sea surface (Bryden et al., 1991; Fang et al., 2003; Li et al., 2011; Zhang et al., 2012),

* Supported by the National Natural Science Foundation of China (Nos. 41406012, 41576060), the Open Fund of State Key Laboratory of Satellite Ocean Environment Dynamics (Second Institute of Oceanography) (No. SOED1613), the Open Fund of Key Laboratory of Global Change and Marine-Atmospheric Chemistry, State Oceanic Administration, China (No. GCMAC1501), and the NSFC-Shandong Joint Fund for Marine Science Research Centers (No. U1406401)
** Corresponding author: qlzhang@qdio.ac.cn

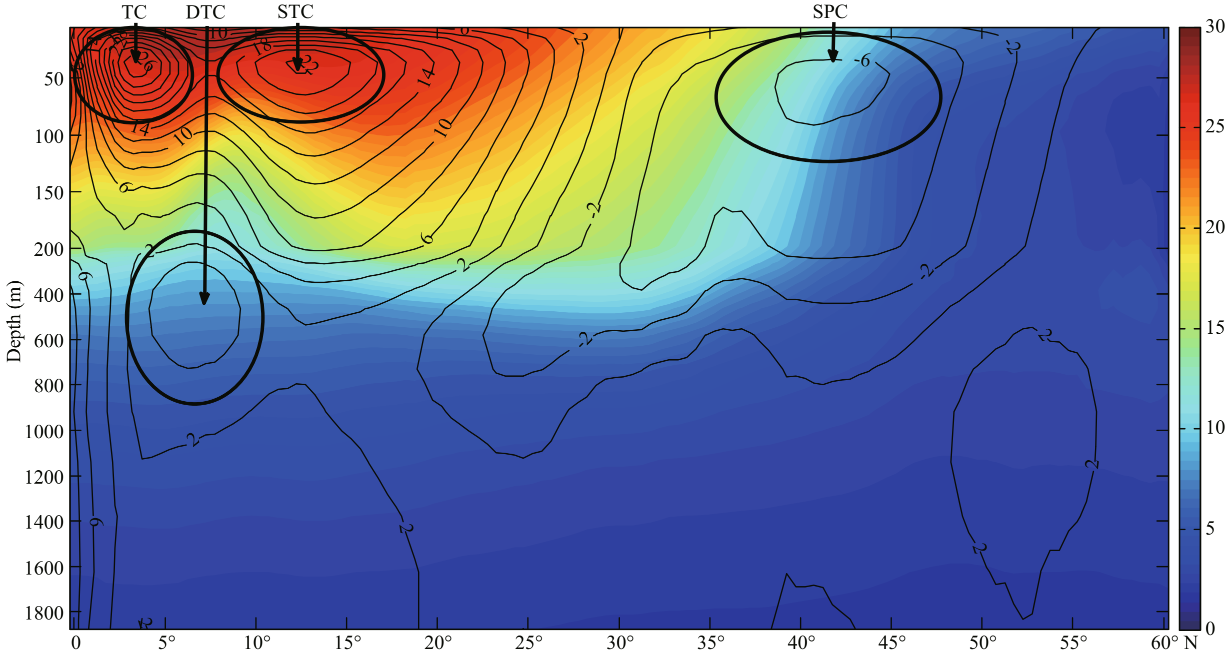

Fig.1 The meridional-vertical distribution of the temperature (shaded), the four cells (contour, 10^6 m^3/s, modified from Fig.1 in Liu et al. (2011)) indicated by the multiyear-averaged meridional stream function in the North Pacific

while the equatorward heat transport is primarily done by the eastern boundary current (such as the California Current) and the thermohaline circulation (including the meridional overturning circulation) in the interior ocean (Rothstein et al., 1998; Wang and Carton, 2002).

It should be noted that the North Pacific Meridional Overturning Circulation (NPMOC) is the general name of all overturning cells in the North Pacific, and it usually includes five cells: the subtropical cell (STC), the tropical cell (TC), the subpolar cell (SPC), the deep tropical cell (DTC) and the thermocline cell (THC) (Bryan, 1991, Liu et al., 1994, 2011; McCreary and Lu, 1994; Lu et al., 1998). As shown in Fig.1, the TC is a strong shallow meridional cell in the low-latitude ocean, whose sea water ascending at the equator flows poleward in the Ekman layer, descends near 5°N and then returns in the thermocline back into the equatorial ocean; the STC is also a shallow meridional cell consisting of subduction of the surface water in the subtropical ocean, equatorward advection of cool subsurface water into the tropical ocean, upwelling at the equator and poleward advection of warm surface water back to mid-latitudes; the DTC is a weaker deep meridional cell between 2°N and 13°N, and the SPC is a weak shallow meridional cell consisting of descending motion over the subtropical ocean and ascending motion over the subpolar region. The THC is a very weak deep meridional cell between 25°N and 65°N (not shown). In contrast to the TC and STC, the DTC is not highly symmetrical, because there is a weak northward flow below 500 m depths between 2°N and 13°N. It is noted that the poleward (equatorward) flow can be named a northward (southward) branch in the present paper.

The STC is a bridge exchanging heat and mass between the tropical and subtropical North Pacific (Pedlosky, 1987; McCreary and Yu, 1992; Liu et al., 1994), and it has a strong effect on the thermohaline structures there, and therefore on the interdecadal variability in the sea surface temperature (SST) in the tropical North Pacific or ENSO (El Nino-Southern Oscillation) (Gu and Philander, 1997; Kleeman et al., 1999; Nonaka et al., 2002; McPhaden and Zhang, 2002; Zhang et al., 2005, 2011), and furthermore on the global climate change (McCreary and Yu, 1992; Liu et al., 1994; Wang and Huang, 2005). In addition, the STC and TC are also important source of water for the Pacific Equatorial Undercurrent in the North Pacific (Lu et al., 1998). Being located between the TC and STC, the DTC will impact directly on the amount of subtropical cold water reaching the equator (Liu et al., 2011). The SPC and THC play an important role in the heat exchange of the North Pacific high latitude regions (Bryan, 1991), and impact the thermohaline structure and climate change between the subtropical and subpolar regions. The NPMOC heat transport is dominated by the overturning cells, while the dominance of meridional overturning is primarily because temperature differences in the ocean are typically much greater within the water column than horizontally. Meridional heat transport due to the thermohaline circulation is dominated by the "overturning component". Thus, the overturning circulation plays a major role in transporting heat. No doubt, these cells should be important channels of meridional heat exchange in the North Pacific. Obviously, it is of important significance to research the North Pacific meridional overturning circulation heat transport for understanding deeply the heat balance of upper North Pacific and climate dynamical mechanisms.

Comparatively speaking, the researches on the NPMOC heat transport have been still few so far, and most previous studies focused the STC poleward heat transport (Klinger and Marotzke, 2000; Riccardo et al., 2014). Based on the numerical results, Klinger and Marotzke (2000) found that the STC poleward heat transport is forced by the wind stress and SST, and the maximum transport is 0.3 PW (1 PW= 1.00×10^{15} W). Moreover, Riccardo et al. (2014) argued that STC decreased over time and its poleward heat transport was reduced, resulting in the positive tropical SST anomalies. However, the STC equatorward heat transport has not been estimated, and the heat transports of other cells (TC, DTC and SPC) have not been examined yet. Moreover, the mechanisms (including dynamical and thermodynamic mechanisms) responsible for the heat transport variability of the cells are not known. Obviously, it is necessary to have a more complete estimation of the NPMOC heat transport on their variations and mechanisms. Therefore, this study is intended to discuss these questions by using the Simple Ocean Data Assimilation (SODA) data. Results of the analysis are presented in the following sections. After a brief description of both the data and methodology in Section 2, the basic characteristics and seasonal changes of the meridional heat transports are analyzed in Section 3. Section 4 examined the dynamical and thermodynamic mechanisms responsible for the seasonal variability in the heat transports, respectively. The results are summarized in Section 5.

2 DATA AND METHOD

2.1 Data description

The results in this study are mainly based on the SODA reanalysis data set from 1958 to 2008 provided by Texas A&M University and the University of Maryland (Carton and Giese, 2008). The 51-year data set (version 2.1.6) is available online in monthly averaged form including meridional and vertical velocities, sea surface height, wind stresses and monthly averages obtained at 0.5°×0.5° horizontal resolution and 26 vertical levels at different vertical spacing, in the upper 1 875.1 m. Since the system assimilated multi-source in situ observations and a series of quality control schemes have already been applied prior to entry into the system, this product is of good comparability with the observed data. Schott et al. (2008) reported that the zonal currents at 140°W on the equator from SODA product are agree well with ADCP (Acoustic Doppler Current Profiler) currents observed during 1990-2005. Song et al. (2014) showed that the temperatures from the SODA product and the XBT (expendable bathythermograph) data in the South China Sea during 1958–2007 are comparable.

The monthly net surface heat flux data are from NCEP (National Centers for Environmental Prediction, USA) reanalysis data. The data set almost covers the whole global (88.542°N–88.542°S, 0°–358.125°E) with Gauss grid (192×94) in the period of 1948 to 2012. And the net sea surface heat flux data in the North Pacific during the period of 1958–2008 is only used in this paper.

2.2 Method

Due to the convergence and divergence in all overturning cells, the northward and southward heat transports in each cell are calculated from

$$Q=\iint C_P\rho(\theta-\theta_0)V\mathrm{d}x\mathrm{d}z,$$

where V is meridional current velocity, ρ is the density of seawater, C_P is the specific heat capacity at constant pressure. Both ρ and C_P are calculated from the temperature, salinity and depth in the text, θ is potential temperature and θ_0 is an arbitrary temperature reference. A meaningful estimate of the meridional ocean heat flux requires the calculation of over a full oceanic section for which there is zero net mass transport; otherwise the calculated heat transport is dependent on an arbitrary temperature reference (Fillenbaum et al., 1997; Johns et al., 2011). In the next text, the θ_0 of TC and STC are reference to the minimum temperature at 200 m depth in their overturning area; the θ_0 of DTC is reference to the minimum temperature at 1 000 m depth, whereas the θ_0 of SPC is reference to the minimum temperature at 500-m depth. It is noted that the reference depth is the

Table 1 Meridional heat transports of each cell

	NHT (PW)	SHT (PW)	Net HT (PW)
TC (3.25°N)	1.75±0.30	-1.69±0.55	0.06±0.50
STC (15.25°N)	0.71±0.65	-0.63±0.53	0.08±1.14
DTC (7.75°N)	0.18±0.03	-0.18±0.11	0.00±0.12
SPC (42.25°N)	0.09±0.05	-0.07±0.09	0.01±0.04

below boundary of each cell and choosing the minimum temperature can avoid negative (positive) contribution in the case of northward (southward) transport. Both the integrations are performed over upper and below boundary depth of northward or southward branch and the western and eastern boundaries of each cell, respectively.

As reported by Liu et al. (2011), the TC is located between 175°E and 100°W shallower 200 m depth with a multiyear-averaged center on Section 3.25°N, the STC between 155°E and 120°W above 200 m depth with a multiyear-averaged center on Section 15.25°N, the DTC between 160°E and 100°W with a multiyear-averaged center on Section 7.75°N, and the SPC between 160°E and 137°W with a multiyear-averaged center on Section 42.25°N. Because the THC is a very weak deep meridional cell, we do not examine its heat transports in the present paper. We calculate mainly the multiyear-averaged northward and southward heat transports of TC, STC, DTC and SPC in their overturning range on Section 3.25°N, 15.25°N, 7.75°N and 42.25°N, respectively. The integral depth varies from cell to cell. The northward heat transport (NHT) of TC and STC is integrated the northward velocity above 50 m, while the southward heat transport (SHT) of TC and STC is integrated the southward velocity above 200 m. The DTC SHT is integrated the southward velocity between 200 m and 600 m, while the DTC NHT is integrated the northward velocity between 600 m and 1 000 m. The SPC SHT is integrated the southward velocity above 75 m, while the SPC NHT is integrated the northward velocity between 75 m and 500 m (Table 1). It is noted that, because the STC center moves meridionally from month to month (the center latitude of other cells basically unchanged) (Liu et al., 2011), we calculate its northward and southward heat transports at the latitude where it centered in each month when we discuss the STC seasonal variation and its mechanism in the text. We choose the STC center latitude as 9.75°N, 9.25°N, 8.75°N, 9.75°N, 10.75°N, 12.25°N, 13.75°N, 16.25°N, 18.75°N, 16.25°N, 13.75 °N and 10.75°N from January to December.

3 RESULT AND DISCUSSION

3.1 The basic feature and seasonal changes of the NPMOC heat transport

As shown in Table 1, among these cells, the TC is strongest with a NHT of (1.75±0.30) PW, a SHT of (-1.69±0.55) PW and a net heat transport of (0.06±0.50) PW, the STC is second with a NHT of (0.71±0.65) PW , SHT of (-0.63±0.53) PW and a net heat transport of (0.08±1.14) PW, the DTC is third with a NHT of (0.18±0.03) PW , SHT of (-0.18±0.11) PW, and a net heat transport of (0.00±0.12) PW, while the SPC is weakest with a NHT of (0.09±0.05) PW, SHT of (-0.07±0.09) PW, and a net heat transport of (0.01±0.04) PW. Note that the STC NHT is a little larger than the numerical results obtained by Klinger and Marotzke (2000) owing to the different data and the different latitude section. It is pointed out that we are concerned about the NHT and SHT and their variation mechanisms, we do not examine the seasonal variation and its mechanism of the net heat transports in the next text.

To examine the seasonal changes of the TC, STC, DTC and SPC heat transports, we calculate their northward and southward heat transports in each month. As shown in Fig.2a and 2b, the TC has strong and quite different seasonal change in its NHT and SHT. The NHT goes through large semiannual oscillation, whose two peaks (valleys) are (2.95±0.47) PW in January ((1.98±0.29) PW in April) and (2.72±0.49) PW in August ((2.28±0.46) PW in October), respectively, while the SHT goes through large annual oscillation with a valley of (-4.24±0.60) PW in October and a peak of (-2.73±0.82) PW in May. It should be pointed out that the annual SHT range is 1.51 PW, while the maximum annual NHT range is 0.97 PW. Obviously, the TC has stronger seasonal change in its SHT than in its NHT.

Different from the TC, the STC NHT and SHT show only anti-phase annual oscillation (Fig.2c, d). The NHT is strongest in March ((2.80±0.36) PW) and weakest in September ((1.04±0.16) PW) with an annual range of 1.76 PW, while the SHT is strongest in September ((-2.70±0.39) PW) and weakest in Apirl ((-1.31±0.29) PW) with an annual range of 1.39 PW. It is also worth noting that the seasonal change in the NHT is stronger than that in the SHT.

The DTC shows quite different seasonal changes in its NHT and SHT (Fig.3a, b). Its NHT is strongest in July ((1.44±0.21) PW) and weakest in March

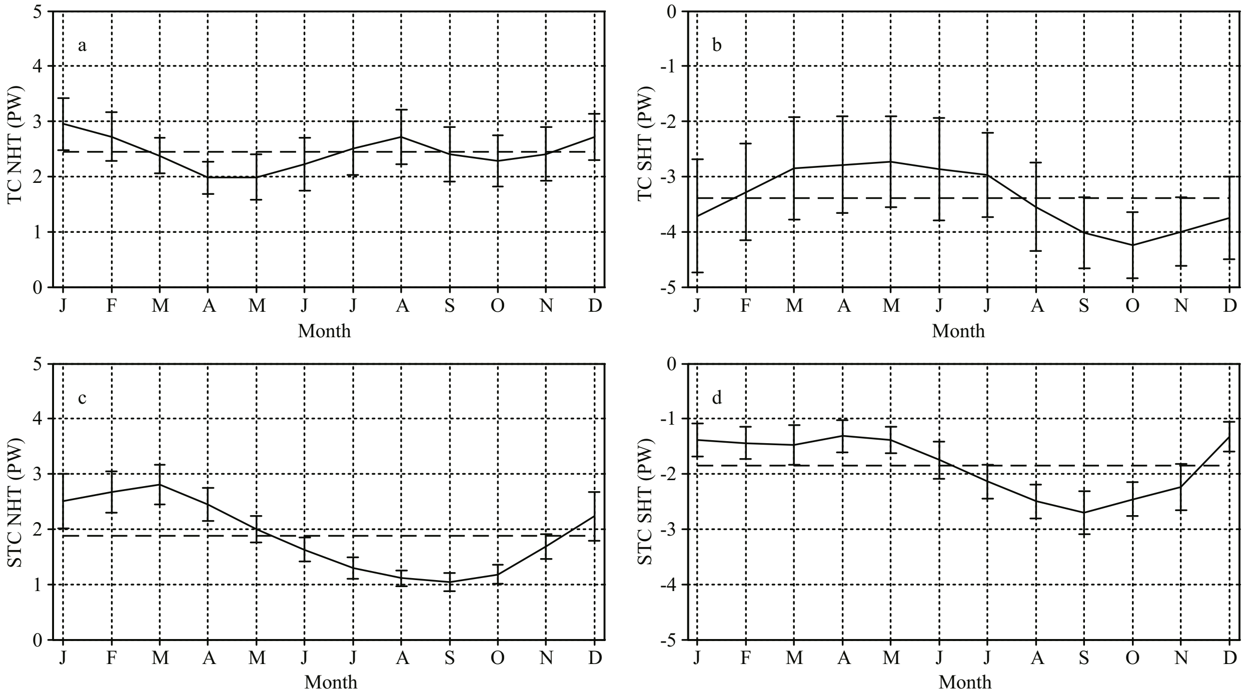

Fig.2 Seasonal variations of the NHT and SHT of TC and STC (unit: PW)

The thick line is the mean cycle of heat transport for each year; the error bar is the standard error of the heat transport from 1958 to 2008; the dotted line is the mean value of the heat transport.

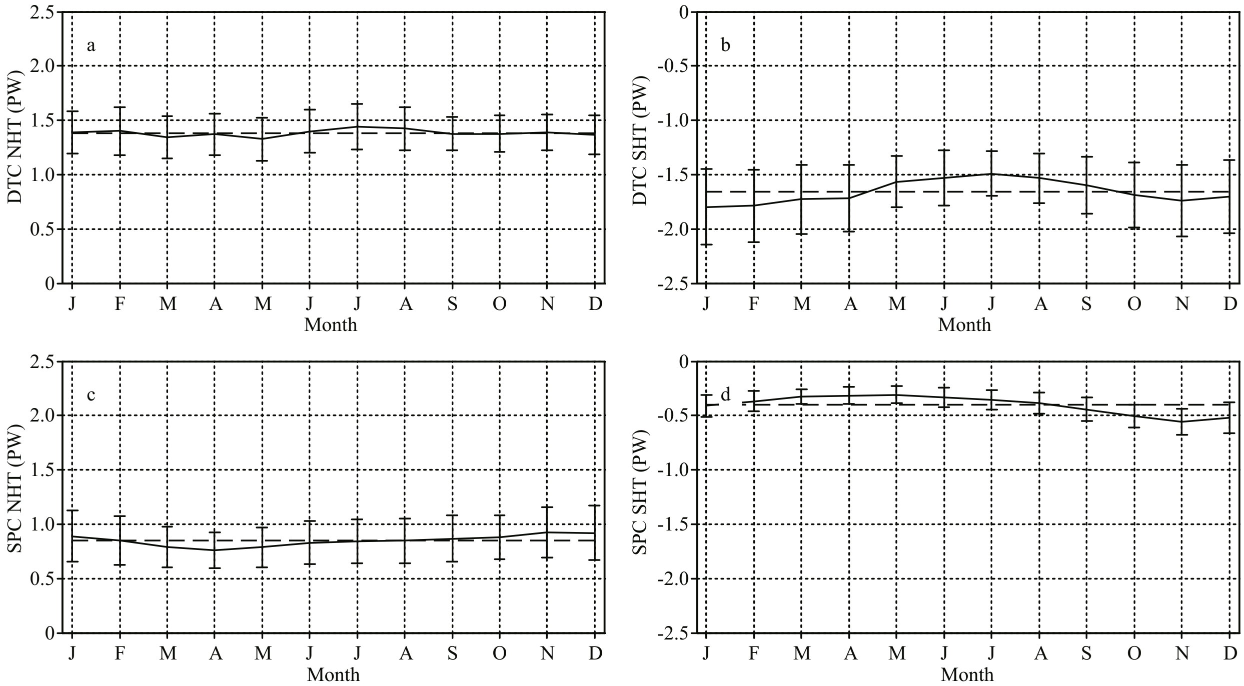

Fig.3 Seasonal variations of the NHT and SHT of DTC and SPC (unit: PW)

The thick line is the mean cycle of heat transport for each year; the error bar is the standard error of the heat transport from 1958 to 2008; the dotted line is the mean value of the heat transport. Noted that the *y* axis has half scale of the Fig.2.

((1.34±0.19) PW) with an annual range of 0.10 PW, while the SHT is strongest in January (-1.80±0.35) PW) and weakest in July ((-1.49±0.21) PW) with an annual range of 0.31 PW. Obviously, the SHT is stronger than the NHT in their seasonal change. Note that the SHT of DTC is out of phase to that of STC, implying that the seasonal change of STC is closely related to that of the DTC.

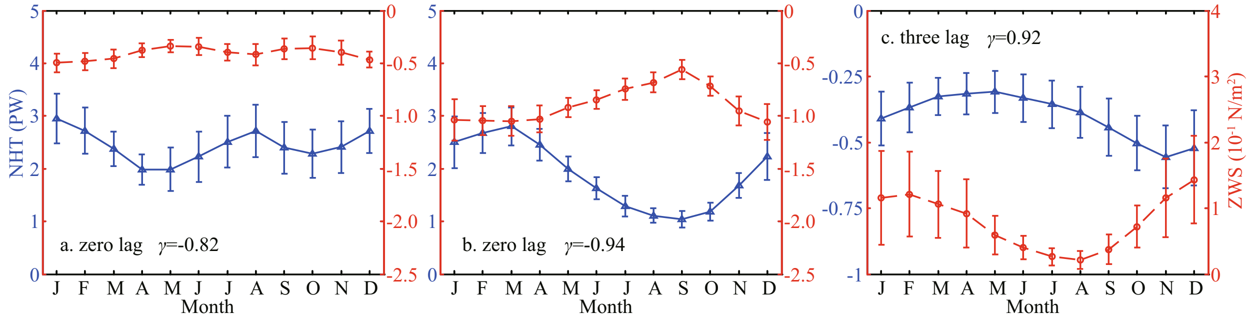

Fig.4 Seasonal changes of zonal wind stress (red line) and meridional heat transports (blue line)

a. TC NHT and zonal wind stress; b. STC NHT and zonal wind stress; c. SPC SHT and zonal wind stress.

The SPC NHT and SHT display in-phase annual oscillation (Fig.3c, d). The NHT is strongest in November ((0.93±0.23) PW) and weakest in April ((0.76±0.17) PW), while the SHT is also strongest in November (-0.56±0.12) PW) and weakest in May ((-0.31±0.08) PW). Because the annual SHT range (0.25 PW) is larger than the annual NHT range (0.17 PW), the seasonal change of the SHT is slightly stronger than that of the NHT.

In the above analysis, the TC and STC are stronger than the DTC and SPC. The four cells have different seasonal changes in their southward and northward heat transports.

3.2 Dynamical and thermodynamic mechanism for seasonal changes of NPMOC heat transport

The four cells of the NPMOC all have significant seasonal changes, but the TC and STC are stronger than the DTC and SPC. Then it raises a question, what process controls these seasonal variations? Previous studies have showed that the STC poleward heat transport is forced by the wind stress and SST (Klinger and Marotzke, 2000), while the zonal wind stress is a main factor driving the seasonal changes in TC and STC volume transports (Liu et al., 2011). In this section, therefore, we examine the impacts of zonal wind stress, thermocline circulations, sea surface heat flux, and sea temperature on the heat transports of the cells respectively to discuss dynamical and thermodynamic mechanisms for seasonal change in NPMOC heat transport.

3.2.1 The zonal wind stress

In order to understand the influence of the zonal wind stress on the seasonal changes in the heat transports of the cells, using the SODA data we calculate the averaged zonal wind stress (the value is averaged between 1° northward and 1° southward latitude of each cell center) over the overturning region of cells in the means of the cycle for each year. In order to convenient discuss, Fig.4 also shows the seasonal changes in the heat transport of the cells.

As shown in Fig.4a, over the TC, the easterly wind prevails throughout year, and has obvious semiannual oscillation. Namely, the local wind stress strengthens in winter and summer and weakens in spring and fall (Fig.4a, red line). The seasonal change in the wind stress is out of phase with that in the TC NHT, and therefore there is close negative relationship between them with a peak coefficient of -0.82 at zero lag (over 99% significant level), implying that the local wind stress is a main factor driving the seasonal change in TC NHT. Namely, when the easterly wind strengthens (weakens), the northward Ekman transport increases (decreases) (Liu et al., 2011), and thus the TC NHT also becomes strong (weak).

Over the STC, the easterly wind also prevails throughout year, as shown in Fig.4b (red line), but zonal wind stress goes through obvious annual oscillation being stronger in winter and weaker in summer. This variation trend is out of phase with that of the STC NHT, and thus close negative relation exists between them with a peak coefficient of -0.94 at zero lag. This shows that the local wind stress could be responsible for the seasonal change of STC NHT. When the easterly wind strengthens (weakens) over the STC area, the northward Ekman transport increases (decreases) (Liu et al., 2011), and thus the STC NHT also becomes strong (weak).

On the contrary, the westerly wind prevails over the SPC center (on Section 42.25°N) throughout year, and displays obvious annual oscillation being strong in winter and weak in summer (Fig.4c). This variation trend is in phase with that of the SPC SHT, and thus there is close correlation (R=0.92) between them when the wind stress leads the SHT for 3 months. This shows that the local wind stress has an important

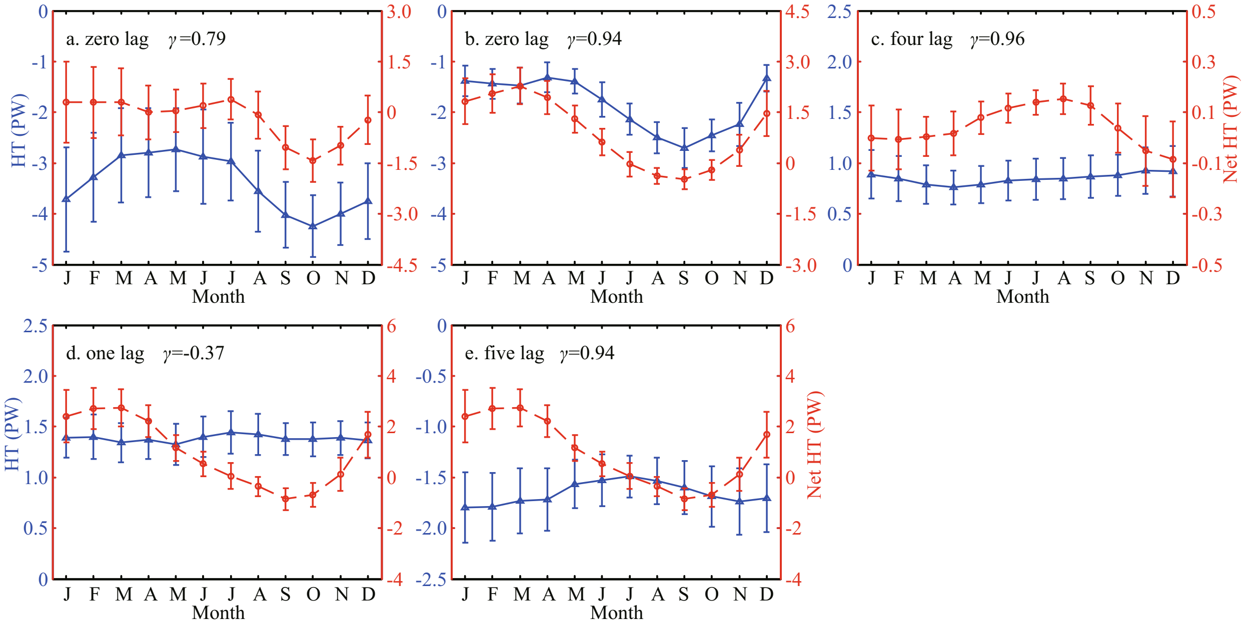

Fig.5 Seasonal changes of the heat transports (blue line) and net heat transports (red line) of the cells

a. TC SHT and net heat transport; b. STC SHT and net heat transport; c. SPC NHT and net heat transport; d. DTC NHT and net heat transport; e. DTC SHT and net heat transport.

influence on the seasonal change of the SPC SHT. Namely, when the westerly wind strengthens (weakens) over the SPC area, the southward Ekman transport increases (decreases) there, and then the SHT also becomes strong (weak) after 3 months.

3.2.2 The thermocline circulations

Because the TC and STC southward branches, the SPC northward branch and DTC northward and southward branches are all located in or below the thermocline (Liu et al., 2011), it is necessary to examine the influences of the ocean thermocline circulations on the seasonal change of their heat transports. For the sake of simplicity, therefore, we take the net heat transports of TC, STC, SPC and DTC (the integrated heat transport above 200 m depth) as their thermocline circulation heat transports in the means of the cycle for each year. The seasonal changes in the net heat transports of the cells are all shown in Fig.5.

It is found from Fig.5a that the TC net heat transport has significant annual oscillation and its trend is consistent with that of the SHT. Thus, there is close positive correlation between them with a peak coefficient of 0.79 at zero lag (over 99% significant level), indicating that the seasonal change in the TC SHT is mostly caused by the thermocline circulation there.

As shown in Fig.5b, the STC net heat transport and SHT have consistent seasonal change each other, and thus close positive correlation exists between them with a peak coefficient of 0.94 at zero lag, showing that the seasonal change of the STC SHT is mainly controlled by the thermocline circulation there.

Similarly, the variation trend of the SPC net heat transport is in phase with that of its NHT (Fig.5c), and thus there is positive correlation (R=0.96, at 99% significant level) between them when the net heat transport leads the NHT for four months. This shows that the thermocline circulation have an important influence on the seasonal change of the SPC NHT.

Figure 5d and e show the seasonal changes of the DTC net heat transport and its NHT and SHT, respectively. The seasonal change in the net heat transport is in phase with that of the DTC SHT, there is positive correlation between them with a peak coefficient of 0.94 when the net heat transport leads the SHT for 5 months. The seasonal change in the net heat transport is poorly related to the DTC NHT. This suggests that the thermocline circulation could be responsible for the seasonal changes of both the DTC SHT. The above analyses suggest that the local zonal wind stress is main dynamical mechanism for the seasonal changes of the TC and STC NHTs and SPC SHT, while the local meridional thermocline circulations are primarily dynamical mechanisms for the seasonal changes of the TC, STC and DTC SHTs and SPC NHT.

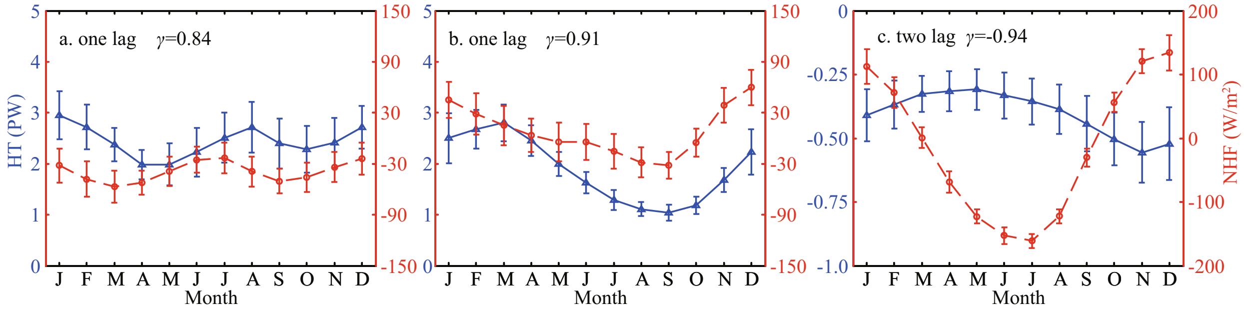

Fig.6 Seasonal changes of net surface heat flux (red line) and meridional heat transport (blue line)

a. TC NHT and the net surface heat flux; b. STC NHT and the net surface heat flux; c. SPC SHT and the net surface heat flux.

3.2.3 The net surface heat flux

Since the thermocline layer gains much less heat from the overlying atmosphere than the surface mixing layer, we focus the impact of net surface heat flux on the TC and STC NHTs and the SPC SHT here at the seasonal time scales. Therefore, using the NCEP surface flux data we calculate the averaged net surface heat flux (the value is averaged between 1° northward and 1° southward latitude of each cell center) on their overturning area in the means of the cycle for each year. The results are shown in Fig.6. In order to convenient discuss, Figure 6 also shows the seasonal changes of the TC and STC NHTs and the SPC SHT, respectively. As shown in Fig.6a, the upper ocean in the TC region releases heat to the overlying atmosphere throughout year, and the net heat flux there goes through semiannual oscillation being larger in March ((-56.99±18.57) W/m^2) and September ((-50.44±14.37) W/m^2) and less in July ((-23.08±17.77) W/m^2) and December ((-24.27±18.82) W/m^2), respectively. This variation trend is consistent with that of the TC NHT, and thus close positive correlation exists between them with a peak coefficient of 0.84 when the net heat flux leads the NHT for 1 month. Namely, when the upper ocean releases less (more) heat 1 month later, the TC NHT would strengthen (weaken). This indicates that the net surface heat flux has important influence on the seasonal change of the TC NHT.

Different from the TC case, the upper ocean in the STC area releases heat to the overlying atmosphere during May–October, but gets heat from the overlying atmosphere in other months (Fig.6b). In the whole year, the obtained heat is maximum ((59.89±20.99) W/m^2) in December, while the lost heat is maximum ((-31.56±15.77) W/m^2) in September. Note that the seasonal change of the net heat flux agrees with that of the STC NHT, and thus close correlation exists between them with a peak coefficient of 0.91 when the net heat flux leads the NHT for 1 month. Namely, when the upper ocean in the STC area gets more heat 1 month later, the STC NHT would strengthen, and vice versa. This indicates that the net surface heat flux has important influence on the seasonal change of the STC NHT.

Similar to the STC case, the upper ocean in the SPC area releases heat to the overlying atmosphere during April-September, but gets heat from the overlying atmosphere in other months (Fig.6c). In the whole year, the obtained heat is maximum ((134.20±27.84) W/m^2) in December, while the lost heat is maximum ((-161.23±11.41) W/m^2) in July. Because the variation trend of the net heat flux is out of phase with that of the SPC SHT, there is close negative correlation between them with a peak coefficient of -0.94 when the net heat flux leads the SHT for 2 months. This implies that the net heat flux has an important impact on the seasonal change of the SPC SHT. Namely, when the obtained heat increases (decreases) 2 months later, the upper ocean becomes warm (cold), and thus the SPC SHT strengthens (weakens).

3.2.4 The sea temperature

In order to understand the influence of the sea temperature on the seasonal changes of the TC and STC SHTs, SPC NHT and the DTC SHT and NHT, we calculate the averaged sea temperature in these cells in the means of the cycle for each year. The results are shown in Fig.7. In order to convenient discuss, Fig.7 also shows the seasonal changes in the corresponding heat transports.

As shown in Fig.7a, the sea temperature in the TC southward branch goes through weak semiannual oscillation with a maximum range of 1.95°C, and this variation trend is out of phase with that of the TC SHT. Thus there is a negative correlation between them with a peak coefficient of -0.85 at zero lag. This

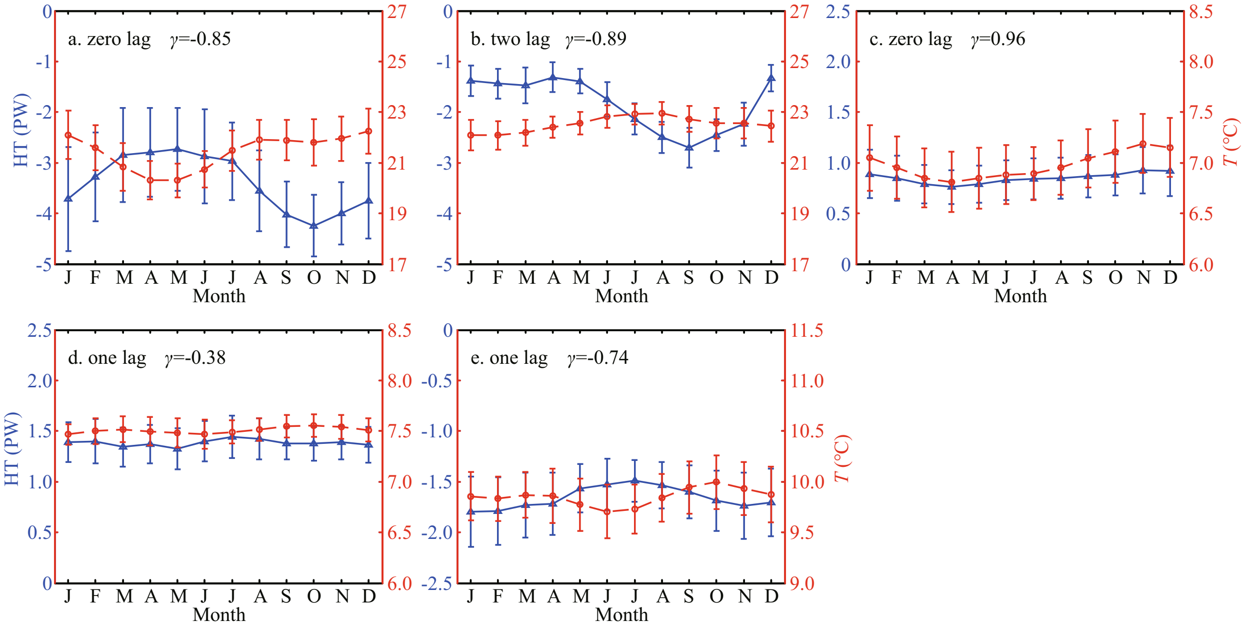

Fig.7 Seasonal changes of the meridional heat transport (blue line) and sea water temperature (red line)

a. TC SHT and sea water temperature; b. STC SHT and sea water temperature; c. SPC NHT and sea water temperature; d. DTC NHT and sea water temperature; e. DTC SHT and sea water temperature.

implies that the sea temperature has influence on the seasonal change of the TC SHT. When the sea temperatures are high (low), the TC SHT would strengthen (weaken).

Different from the TC case, the sea temperature in the STC southward branch displays strong annual oscillation with a range of 0.88°C, and this variation trend is out of phase with that of the STC SHT (Fig.7b), and thus close negative correlation exists between them with a peak coefficient of -0.89 when the temperature leads the SHT for 2 month. This indicates that the sea temperature makes large contribution to the seasonal change of the STC SHT. When the sea temperatures are high (low), the STC SHT would strengthen (weaken).

Although the sea temperature of the SPC northward branch has weak seasonal change with a range of 0.38°C, as shown in Fig.7c, its variation trend agrees well with that of the SPC NHT, and thus close positive correlation exists between them with a peak coefficient of 0.96 at zero lag, showing that the sea temperature has larger influence on the seasonal change of the SPC NHT. When the sea temperatures are high (low), the SPC NHT would strengthen (weaken).

Figure 7d and e shows the seasonal changes of the sea temperature in the DTC northward and southward branches and their corresponding heat transports. The sea temperature in DTC southward branch has stronger seasonal change (with an annual range of 0.29°C) and its variation trend is out of phase with that of the DTC SHT, and thus there is negative correlation between them with a peak coefficient of -0.74 when the temperature leads the SHT for 1 month. The sea temperature in the DTC northward branch has weak seasonal change with a maximum range of 0.09°C, and it is poorly related to the DTC NHT. The result shows that the sea temperature of the DTC southward branch has influence on the DTC SHT.

The above analyses suggest that the net surface heat flux is main thermodynamic mechanism for the seasonal changes of the TC and STC NHTs and SPC SHT, while the local sea temperature is primarily thermodynamic mechanism for the seasonal changes of the TC, STC and DTC SHTs and SPC NHT.

4 CONCLUSION

Based on the SODA data from 1958 to 2008, we investigated the basic characteristics and seasonal changes of the southward and northward heat transports of TC, STC, DTC and SPC centered in their overturning range, respectively. In addition, we also examined the dynamical and thermodynamic mechanisms responsible for these heat transports at the seasonal time scales. The results are summarized as follows.

1. Among the four cells, the TC is strongest with a NHT of (1.75±0.30) PW and SHT of (-1.69±0.55) PW, the STC is second with a NHT of (0.71±0.65) PW and SHT of (-0.63±0.53) PW, the DTC is third with a

NHT of (0.18±0.03) PW and SHT of (-0.18±0.11) PW, while the SPC is weakest with a NHT of (0.09±0.05) PW and SHT of (-0.07±0.09) PW.

2. The four cells have different seasonal changes in their southward and northward heat transports. Of all, the TC NHT (SHT) is strongest in January (October) and weakest in April (May), the STC NHT (SHT) is strongest in March (September) and weakest in September (Apirl), the DTC NHT (SHT) is strongest in July (January) and weakest in March (July), while the SPC NHT (SHT) is strongest in November (November) and weakest in April (May). Comparatively speaking, the TC has larger variation amplitude in its SHT than in its NHT, so do both the DTC and SPC, but the STC has less variation amplitude in its SHT than in its NHT.

3. Since the four cells have different seasonal changes in their southward and northward heat transports, their dynamical mechanisms are also different each other. Of all, the local zonal wind stress is responsible for the seasonal changes in the TC and STC NHTs and SPC SHT, while the local thermocline circulations are for the seasonal changes of the TC, STC and DTC SHTs and SPC NHT.

4. Similarly, the thermodynamic mechanisms for the NHT and SHT of the cells are also different each other. The net surface heat flux is responsible for the seasonal changes in the TC and STC NHTs and SPC SHT, while the local sea temperatures are for the seasonal changes of the TC, STC and DTC SHTs and SPC NHT.

5 ACKNOWLEDGMENT

The authors duly acknowledge the various data sources for the freely available data: SODA data from http://apdrc.soest.hawaii.edu/dods/public_data/SODA/soda_pop2.1.6, NCEP reanalysis data from https://www.esrl.noaa.gov/psd/data/gridded/data.ncep.reanalysis.derived.surfaceflux.html.

References

Bryan K. 1991. Poleward heat transport in the ocean: A review of a hierarchy of models of increasing resolution. *Tellus A*: *Dynamic Meteorology and Oceanography*, **43**(4): 104-115.

Bryden H L, Roemmich D H, Church J A. 1991. Ocean heat transport across 24°N in the Pacific. *Deep Sea Research Part A*. *Oceanographic Research Papers*, **38**(3): 297-324.

Carton J A, Giese B S. 2008. A reanalysis of ocean climate using Simple Ocean Data Assimilation (SODA). *Monthly Weather Review*, **136**(8): 2 999-3 017.

Fang G H, Wei Z X, Choi B H, Wang K, Fang Y, Li W. 2003. Interbasin freshwater, heat and salt transport through the boundaries of the East and South China Seas from a variable-grid global ocean circulation model. *Science in China Series D*: *Earth Sciences*, **46**(2): 149-161.

Fillenbaum E R, Lee T L, Johns W E, Zantopo R J. 1997. Meridional heat transport variability at 26.5°N in the North Atlantic. *Journal of Physical Oceanography*, **27**(1): 153-174, https://doi.org/10.1175/1520-0485(1997)027<0153:MHTVAN>2.0.CO;2.

Gu D F, Philander S G H. 1997. Interdecadal climate fluctuations that depend on exchanges between the tropics and extratropics. *Science*, **275**(5301): 805-807.

Hsiung J. 1985. Estimates of global oceanic meridional heat-transport. *Journal of Physical Oceanography*, **15**(11): 1 405-1 413.

Johns W E, Baringer M O, Beal L M, Cunningham S A, Kanzow T, Bryden H L, Hirschi J J M, Marotzke J, Meinen C S, Shaw B, Curry R. 2011. Continuous, array-based estimates of atlantic ocean heat transport at 26.5°N. *Journal of Climate*, **24**(10): 2 429-2 449.

Kleeman R, McCreary J P, Klinger B A. 1999. A mechanism for generating ENSO decadal variability. *Geophysical Research Letters*, **26**(12): 1 743-1 746.

Klinger B A, Marotzke J. 2000. Meridional heat transport by the subtropical cell. *Journal of Physical Oceanography*, **30**(4): 696-705.

Li P, Zhang Q L, Liu H W, Xu J P. 2011. Seasonal variation of the North Pacific meridional net heat transport. *Advances in Marine Science*, **29**(3): 275-284. (in Chinese with English abstract)

Liu H W, Zhang Q L, Duan Y L, Hou Y J. 2011. The three-dimensional structure and seasonal variation of the North Pacific meridional overturning circulation. *Acta Oceanologica Sinica*, **30**(3): 33-42.

Liu Z Y, Philander S G H, Pacanowski R C. 1994. A GCM study of tropical-subtropical upper-ocean water exchange. *Journal of Physical Oceanography*, **24**(12): 2 606-2 623.

Lu P, Mccreary J P, Klinger B A. 1998. Meridional circulation cells and the source waters of the Pacific equatorial undercurrent. *Journal of Physical Oceanography*, **28**(1): 62-84.

McCreary J P, Lu P. 1994. Interaction between the subtropical and equatorial ocean circulations: the subtropical cell. *Journal of Physical Oceanography*, **24**(2): 466-497.

McCreary J P, Yu Z J. 1992. Equatorial dynamics in a 212-layer model. *Progress in Oceanography*, **29**(1): 61-132.

McPhaden M J, Zhang D X. 2002. Slowdown of the meridional overturning circulation in the upper Pacific Ocean. *Nature*, **415**(6872): 603-608.

Nonaka M, Xie S P, McCreary J P. 2002. Decadal variations in the subtropical cells and equatorial pacific SST. *Geophysical Research Letters*, **29**(7): 20-1-20-4, https://doi.org/10.1029/2001GL013717.

Pedlosky J. 1987. An inertial theory of the equatorial undercurrent. *Journal of Physical Oceanography*, **17**(11): 1 978-1 985.

Riccardo F, Molteni F, Kucharski F. 2014. Pacific interdecadal variability driven by tropical-extratropical interactions. *Climate Dynamics*, **42**(11-12): 3 337-3 355.

Rothstein L M, Zhang R H, Busalacchi A J, Chen D K. 1998. A numerical simulation of the mean water Pathways in the subtropical and tropical Pacific Ocean. *Journal of Physical Oceanography*, **28**(2): 322-343.

Schott F A, Stramma L, Wang W Q, Giese B S, Zantopp R. 2008. Pacific subtropical cell variability in the SODA 2.0.2/3 assimilation. *Geophysical Research Letters*, **35**(10): L10607, https://doi.org/10.1029/2008GL033757.

Song W, Lan J, Liu Q Y, Sui D D, Zeng L L, Wang D X. 2014. Decadal variability of heat content in the South China Sea inferred from observation data and an ocean data assimilation product. *Ocean Science*, **10**(1): 135-139.

Trenberth K E, Solomon A. 1994. The global heat-balance: heat transports in the atmosphere and ocean. *Climate Dynamics*, **10**(3): 107-134, https://doi.org/10.1007/BF00210625.

Wang J D, Carton J A. 2002. Seasonal heat budgets of the north Pacific and north Atlantic Oceans. *Journal of Physical Oceanography*, **32**(12): 3 474-3 489.

Wang Q, Huang R X. 2005. Decadal variability of pycnocline flows from the subtropical to the Equatorial Pacific. *Journal of Physical Oceanography*, **35**(10): 1 861-1 875.

Zhang L P, Wu L X, Yu L S. 2011. Oceanic origin of a recent La Niña-like trend in the Tropical Pacific. *Advances in Atmospheric Sciences*, **28**(5): 1 109-1 117.

Zhang Q L, Hou Y J, Yan T Z. 2012. Inter-annual and inter-decadal variability of Kuroshio heat transport in the East China Sea. *International journal of Climatology*, **32**(4): 481-488.

Zhang Q, Yang H J, Zhong Y F, Wang D X. 2005. An idealized study of the impact of extratropical climate change on El Niño-Southern Oscillation. *Climate Dynamics*, **25**(7-8): 869-880.