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Freshening of the intermediate water of the South China Sea between the 1960s and the 1980s*

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Abstract Variation in intermediate water salinity in the South China Sea (SCS) between the 1960s and 1980s was studied using historical hydrographic data. The results demonstrate that the water was significantly fresher in the 1980s than in the 1960s, indicating that vertical mixing at intermediate water depth was reduced in the 1980s. This was partially because of the change of the SCS meridional overturning circulation (MOC) connecting local intermediate water with deep water. Data assimilation showed a 0.5 Sv (1 Sv=10^6 m^3/s) reduction in the strength of the MOC, which is about one third of the mean SCS MOC. Because the SCS MOC is linked to the Pacific Ocean, such an interdecadal variation in the intermediate water SCS may reflect anthropogenic climate change in the world ocean.

Keyword: decadal variability; intermediate water; meridional overturning circulation; South China Sea (SCS)

1 INTRODUCTION

The South China Sea (SCS) is a large, semi-enclosed marginal sea bounded by the Indo-China Peninsula to the west, China and Taiwan Island to the north, the Philippine Islands to the East, and Borneo Island to the South. It connects the East China Sea, the Pacific Ocean, the Sulu Sea, the Java Sea and the Indian Ocean via several straits, of which the Luzon Strait is the deepest (>2 500 m). The water exchange between the SCS and the Pacific through the Luzon Strait is persistent. Studies have indicated that the lower temperature and salinity of the intermediate water and the higher density of cold deep water in the SCS comes mainly from the Pacific Ocean (Broecker et al., 1986; Qu et al., 2000, 2006).

The mean meridional circulation in the South China Sea is characterized by the outflow of the intermediate water mass via a deep channel of the Luzon Strait (Li and Qu, 2006). Since there is no local origin for deep water in the SCS, in any season, the Luzon Strait exchange below 1 500 m is the only source.

The response of the ocean to global warming has become a core of oceanographic study (Levitus et al., 2001). Reponses can occur in both the world ocean and within regional basins. By analyzing the existing multi-decadal records of surface and subsurface temperature and salinity data, with respect to spatial and temporal variability, regional waters' responses have been reported (Rahmstorf, 1995; Wong et al., 1999; Kim et al., 2001; Gille, 2002; Auad et al., 2003).

A crucial question connected with oceanic response to global warming is the reliable detection of a systematic trend in water mass properties, on the time scale of decades to a century. This detection is rendered difficult by the presence of substantial variability on time scales both shorter and longer

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than the one of interest. Greenhouse gases have increased rapidly over the last century, e.g., carbon dioxide (CO_2) by about 25%. Documented oceanic change over the same time period is at a much smaller magnitude; e.g., the East (Japan) Sea has been in a warming trend during the past 40 plus years, with 0.1–0.5°C of warming in the upper 1 000 m (Kim et al., 2001). This is smaller than climate models would suggest and marred by non-negligible uncertainties.

Based on the available historic data, changes in the intermediate water of the SCS between two typical periods: 1960s (Phase A) and 1980s (Phase B), was examined. During the two phases, several cruises were carried out, resulting in many temperature and salinity profiles; moreover, the global surface air temperature anomaly in the 1980s is much greater than that in 1960s (Fig.1a), and the sea surface temperature (SST) of the whole SCS in the 1980s was higher than that of the 1960s as well. The difference of the averaged SST during these two periods exceeds 0.1°C.

2 DATA AND METHODS

The data used for analysis of the SCS intermediate water (SCSIW) during the 1960s and 1980s come from the in-situ deep casts obtained from many cruises of the South China Sea Institute of Oceanology (SCSIO), most of which reached 2 000-m depth or deeper. Between 1960s and 1980s, more than 100 hydrographic profiles were surveyed to the highest standards then available. These cruises were conducted by the SCSIO staff or through domestic and international joint research programs.

The procedure for quality control of the data for this study included the removal of obviously erroneous records (e.g., temperature higher than 8°C below 1 000 m or salinity lower than 30 below 100 m) and of profiles with measurements only in the shallow (<100 m) waters and at depth (>700 m) without any data in between (Qu et al., 1999). Figure 1b shows the distribution of the stations with data from depths below 2 000 m. The temperature and salinity of the two phases are averaged on isopycnal surfaces in order to avoid artificial mixing owing to false interpolation.

In order to interpret the change between the two phases, the SCS MOC strength and average wind-energy input into the surface current were calculated for each phase using velocity and wind-stress data from the SODA 1.4.2 (Simple Ocean Data Assimilation) (Carton and Giese, 2008). The velocity data from SODA had an average 0.5°×0.5° horizontal resolution and 40 vertical levels with 10 m spacing near the surface. The calculation of wind-energy input into the surface current was based on the scalar product of wind stress and surface current and surface geostrophic current (Huang et al., 2006). The meridional overturning stream-function has been extensively used to describe the MOC. In comparison with the velocity field, it provides a better representation of the meridional mass transport of ocean circulation.

The definition of the meridional overturning stream-function is as follows. In the spherical coordinate, zonal integration of the continuity equation of the sea water from the western boundary

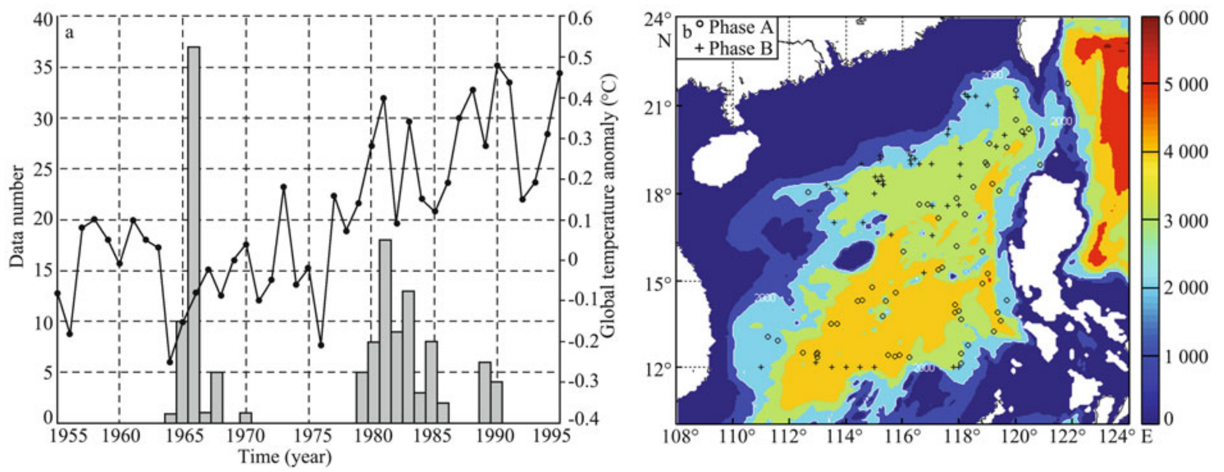

Fig.1 a. Numbers of temperature and salinity profiles obtained during Phase A (1960s) and Phase B (1980s) (histogram), superimposed with the global surface air temperature anomaly (curve); b. Distribution of deep stations of Phase A and B in the northern South China Sea (marked by circles and crosses, respectively)

λ_w to the eastern boundary λ_e gives

$$\int_{\lambda_w}^{\lambda_e} \frac{\partial(v \cdot \sin\theta)}{\partial\theta} d\lambda + a \cdot \sin\theta \int_{\lambda_w}^{\lambda_e} \frac{\partial w}{\partial z} d\lambda = 0, \quad (1)$$

As Eq.1 is non-divergent, we may define the meridional overturning stream-function ψ by

$$\int_{\lambda_w}^{\lambda_e} v \cdot \sin\theta d\lambda = -\frac{\partial\psi}{\partial z}, \quad (2)$$

$$\int_{\lambda_w}^{\lambda_e} w \cdot a \cdot \sin\theta d\lambda = -\frac{\partial\psi}{\partial\theta}. \quad (3)$$

Eq.2 is used to calculate the SCS meridional overturning stream-function in this paper. The zonal integration is performed across the SCS basin, the resulted stream-function corresponds to the SCS meridional transport.

3 RESULT

3.1 SCSIW in the 1960s and 1980s

The SCSIW is located at about 400–800 m, and the water mainly outflows through the Luzon Strait and modulates the North Pacific Intermediate Water (NPIW). The temperature and salinity of the SCSIW is much higher than the NPIW that is defined by a salinity minimum in the density range of 26.5–27.0 σ_θ (Qu et al., 2000). In the salinity minimum layer, salinity is found to be lowest (<34.25) east of the Luzon Strait and the minimum salinity of the SCSIW reaches a value of 34.39. The SCS Deep Water is located at about 1 000–2 500 m. In the deep layer, the properties of the SCS Deep Water are similar to those of the North Pacific Deep Water; the salinity of the SCS Deep Water is a little lower than that of the North Pacific Deep Water and the temperature of the SCS Deep Water is a little higher than that of the North Pacific Deep Water (Fig.2).

In the interior of the SCS, a prominent feature is that both salinity maxima and minima can be seen in most parts of the SCS. The depth range of intruded North Pacific Tropic Water (NPTW) in the SCS is between 120 and 150 m. The depth of the intruded NPIW in the SCS is located between 480 and 500 m. In the SCS, the maximum salinity is lower than 34.6 in the salinity maximum layer, and the minimum salinity is between 34.4 and 34.5 in the salinity minimum layer.

The total pairs of the temperature-salinity profiles

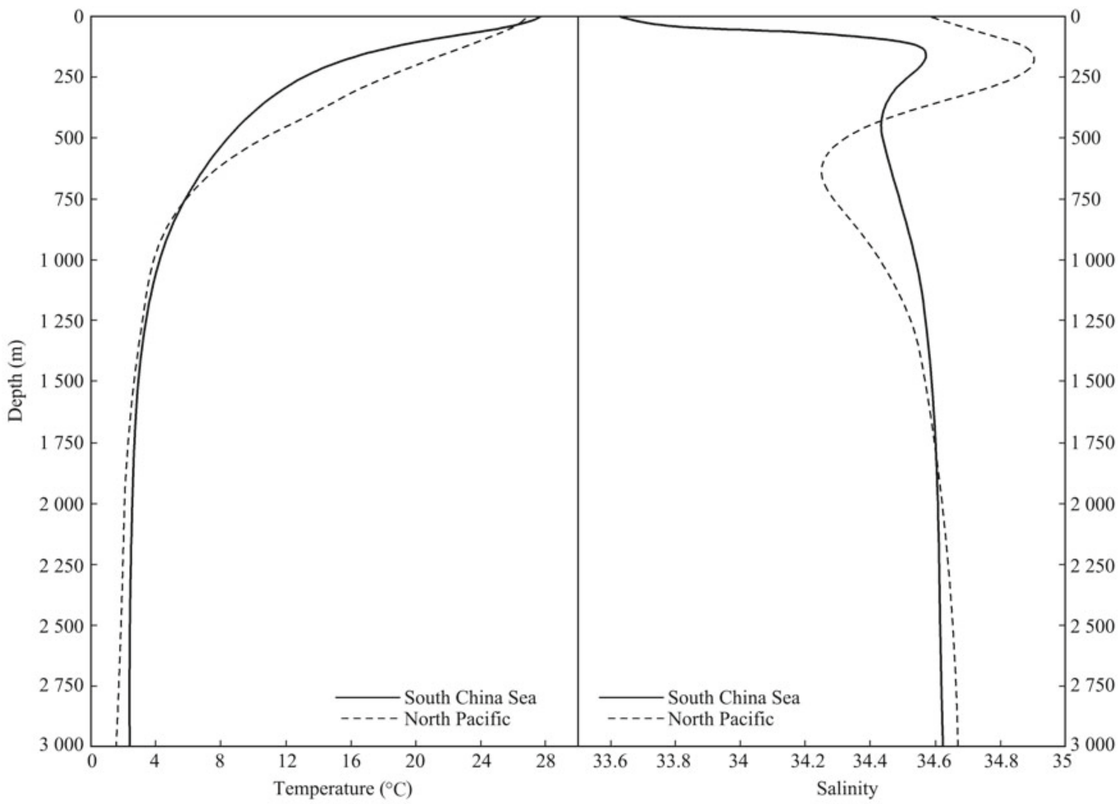

Fig.2 The vertical distribution of mean temperature and salinity in the SCS (Solid line) and North Pacific (dashed line)

The data are averaged temperature and salinity of profile WOD01.

during Phases A and B are 743 and 659, respectively, and each variable was averaged along isopycnal surfaces. The sum of the squares of the difference between the best fit salinity and original salinity was 0.996 9 during Phase A and 0.616 9 during Phase B; the mean distance between the best fit point and the original *T*-*S* scatter point was 0.020 2 during Phase A and 0.019 1 during Phase B.

The scattered *T*-*S* diagram of Phases A and B and the best fit lines of these two phases (Fig.3) indicates that the salinity of the SCSIW during the 1980s was about 0.05 lower than that during the 1960s, while the global surface air temperature was much higher during the 1980s than that during the 1960s (Fig.1a).

3.2 Spin-down of model-based MOC

One of the major elements of today's ocean system is a conveyor-like Atlantic MOC that includes the northward transport of warm, salty tropical and subtropical surface water and the southward transport of cold North Atlantic Deep Water. In the SCS, the MOC is also remarkable and drives the water transport inside the SCS (Wang et al., 2004). But the SCS MOC oscillates in time.

The averaged SCS MOCs during these two phases are shown in Fig.4a and 4b. North of 18°N is dominated by Luzon Strait (open area) transport, thus 18°N was used as a cut-off point in Fig.4. The basic structure of the MOCs during the two phases is same. The shape of the shallow SCS MOC is similar to the results of a SCS regional model (Wang et al., 2004). Interestingly, the strength of the averaged shallow (above 800 m) SCS MOC during the 1980s is much more powerful than that of the 1960s, while the deep SCS MOC is weaker during the 1980s than that of the 1960s (Fig.4c). This demonstrates that the water's mean southward movement in the intermediate layer is much stronger during the 1980s relative to that in the 1960s, which enhances the intrusion of the NPIW into the SCS through the Luzon Strait during the 1980s compared to the 1960s. Extensive water exchange between the SCS and the Pacific Ocean through the Luzon Strait exists all year round (Qu, 2002). Several studies demonstrate that the Luzon Strait transport has a large effect on the SCS circulation (Yuan, 2002; Qu et al., 2004).

The salinity of the NPIW is much lower than that of the SCSIW (Li and Qu, 2006). Owing to the effect of mixing processes, the SCSIW became much fresher during the 1980s than in the 1960s. This can partly explain the phenomenon in Fig.3. The averaged deep

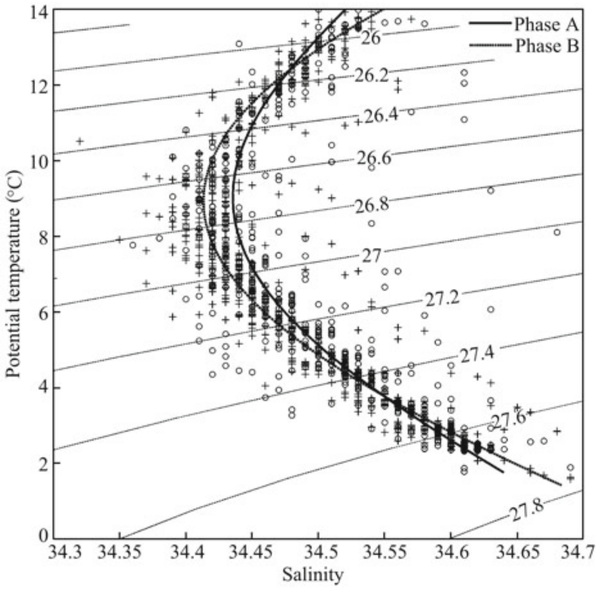

Fig.3 *T-S* scatter diagram of Phase A (1960–1969; circles) and Phase B (1980–1989; crosses), and the best fit lines of these two phases

SCS MOC during the 1980s is weaker than that in the 1960s, which suggests that the renewal speed of the SCS Deep Water may have become slower during the 1980s relative to that in the 1960s. This phenomenon is another possible cause of the freshening of SCSIW.

4 DISCUSSION

The above results indicate that between the 1960's and 1980's the formation of intermediate water in the SCS became faster which may have played a leading role in the freshening of SCSIW. At the same time deep water formation became slower, even shifting from deep water formation to intermediate water formation in the SCS, the thermohaline conveyor of the SCS may have gradually shifted from the 1960s' mode (Fig.4d) to the 1980s' mode where the renewal of the SCS deep water became slower and may have even stopped, again stimulating the freshening of SCSIW. This phenomenon has taken place in the East (Japan) Sea, and the temporal change of the Japan Sea circulation has been related to global climate changes and thus anthropogenic environmental changes (Gamo et al., 2001). In addition, recent observational and theoretical works have shown a decrease of bottom water formation in the northern Atlantic Ocean, possibly because of an increase in anthropogenic greenhouse gases (Wood et al., 1999).

Wind-stress energy input into surface geostrophic currents is closely related to gravitational potential

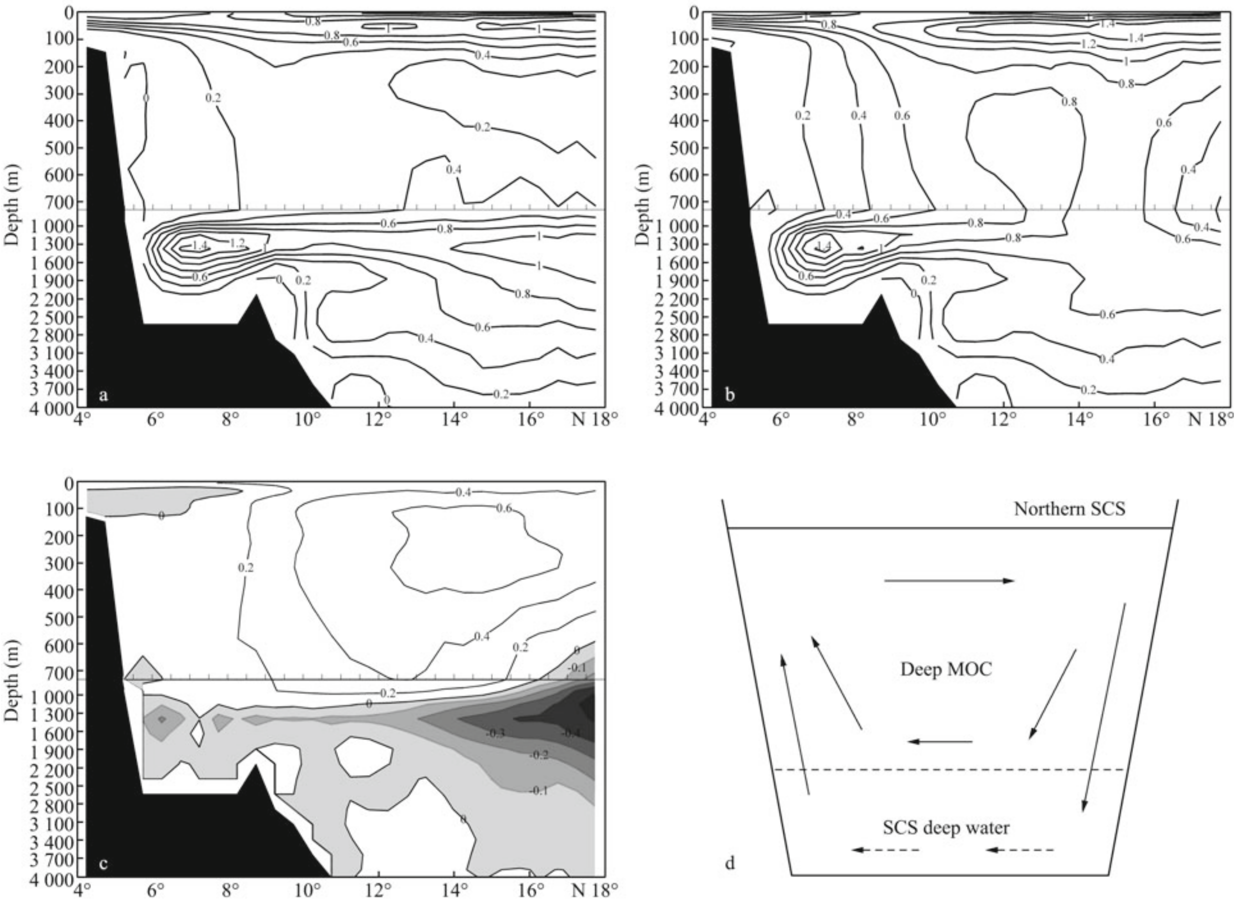

Fig.4 Averaged Meridional Overturning Circulation (MOC) during the 1960s (a) and 1980s (b), respectively; c. Difference of the averaged MOC (1980s minus 1960s); d. Schematic map of the meridional circulation in the SCS during Phase A (1960s)

energy in the ocean, and this energy supports the maintenance of the SCS MOC in the ocean. The change of the SCSIW may have a close relationship with the decadal variability of wind-energy input to the SCS (Fig.5). The picture shows that the wind-energy input to the SCS is larger in the Luzon Strait and off the Vietnam coast, which has led to the greater intrusion of the NPIW into the SCS.

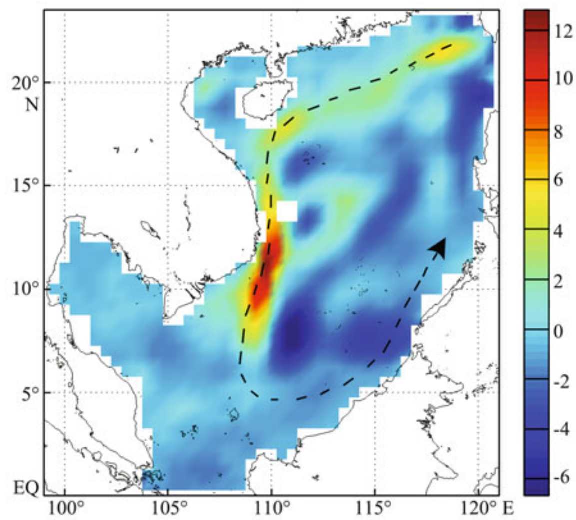

Fig.5 Difference of wind-energy input into the surface current in the SCS between the mean during Phase B (1980s) and the mean during Phase A (1960s)

5 CONCLUSION

The results of the study show that the SCSIW became fresher during the 1980s relative to the 1960s, which can be explained by the decadal variation of shallow meridional overturning circulation in the South China Sea. The increased shallow SCS MOC and related increase in NPIW inflow is the leading cause of freshening of the SCSIW. The slowing down of deep SCS MOC is another possible cause. The South China Sea is a typical example where climatological change might have impacts on deep meridional overturning circulation. Thus, the South China Sea is one of the important seas for monitoring regional and global climate change, and more comprehensive observations should be carried out in the future.

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