Impact of climate variability of the Western Tropical Pacific on maximum salinity water in the South China Sea

Joachim W. Dippner · Sarah C. Weber · Ajit Subramaniam

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Abstract
Salinity observations in the Vietnamese upwelling area in June 2016 indicated a significant increase in the salinity of the maximum salinity water (MSW). The source of MSW inflow into the South China Sea (SCS) is a mixture of the Western North Pacific Central Water and the North Pacific Equatorial Water. Although the East Asian winter monsoon is correlated with both the El Niño Southern Oscillation (ENSO) and the Pacific Decadal Oscillation (PDO), the mean salinity of MSW is only spuriously lag correlated to the PDO, but highly correlated to all tropical climate modes (except El Niño Modoki) with a time lag up to 7 months. Composite analyses indicate that the modulation of ENSO by a PDO in a positive phase results in optimal inflow conditions. A comparison of two post-El Niño years with different PDO polarity (negative in 2003 and positive in 2016) shows that the dominant driver is the variability in outgoing long-wave radiation (OLR) and in zonal wind in the tropics. In 2003, enhanced convective activity over the West Pacific warm pool resulted in a cyclonic circulation. In 2016, convective activity was weak and an anticyclonic circulation was intensified, which transported the saltier North Pacific Equatorial Water into the SCS. This observed increase in the salinity of MSW requires a modification of the previous definitions of characteristic water masses, which is presented here. The question of whether or not the increase in MSW salinity is a transient phenomenon cannot be answered. It might be possible that the increase in salinity is related to global warming.

Keywords ENSO · PDO · East Asian monsoon · Water mass analysis · South China Sea · Western Tropical Pacific

1 Introduction

The Luzon Strait Transport (LST) is a part of the South China Sea throughflow (SCSTF), which is an important conveyor belt between the Pacific and Indian Oceans (Qu et al. 2006). The LST is driven by two different water masses entering the South China Sea (SCS) through Luzon Strait: the Western North Pacific Central Water (WNPCW) and the North Pacific Equatorial Water (NPEW, Figs. 1 and 2). Both water masses have distinct salinities (Tomczak and Godfrey 1994) and form the maximum salinity water (MSW), which is a characteristic water mass in the SCS (Rojana–Anawat et al. 2001) and the core water of the seasonal upwelling off the Vietnamese coast (Dippner et al. 2007). Rojana–Anawat et al. (2001) have previously identified seven characteristic water masses in the western SCS, which were later improved by Dippner and Loick–Wilde 2011. Recent observations in the upwelling area of SCS (Figs. 2) by the “RV Falkor” in June 2016 indicated a statistically significant increase in the salinity of MSW. The consequence was that a subset of the hydrographic observations did not fit into the definition of characteristic water masses given by Dippner and Loick–Wilde (2011). As such, the water mass definition needs a modification, which is presented here. The scientific questions considered here are why did the MSW salinity increase and what is the mechanism behind this phenomenon?

The circulation in the SCS (e.g., Dale 1956; Wyrski 1961; Pohlmann 1987; Metzger and Hurlburt 1996) is forced by a spatial homogeneous northeast monsoon (NEM), during winter and by the curl of the wind stress during the southwest monsoon (SWM) in summer (Chao et al. 1996). The
circulation in the southern SCS (Fig. 1) is characterized by a cyclonic gyre during winter and an anticyclonic gyre during summer. In the northern part of the SCS, the circulation is forced by the monsoon and the LST (Hu et al. 2000) forming a basin wide cyclonic gyre in the SCS during winter. In summer (dashed line in Fig. 1), both gyres form a dipole circulation cell causing an offshore current off the coast of Vietnam at ~12°N (Dippner et al. 2007).

The LST, which represents the leaking path of the Kuroshio intrusion into SCS (Sheremet 2001; Nan et al. 2015), occurs all year round with a pronounced semiannual signal, with greater strength during winter and summer monsoon and weaker strength during inter-monsoon (Qu et al. 2000). A process analysis indicates that ~95% of the salinity variability in the LST is caused by advection due to the Kuroshio intrusion; the impact of river discharge and the air–sea freshwater flux are much smaller (Nan et al. 2016). Hence, the reason for the variability of salinity is located in the dynamically complex Western Tropical Pacific (WTP; Qu et al. 2004; Yan et al. 2013; Nan et al. 2015 and references therein; Kidwell et al. 2017; Hu et al. 2017; Roxy et al. 2019).

Besides the seasonal variability, the dynamics of WTP and adjacent seas are driven by different multi-time scale processes. On an intra-annual time scale, two frequency bands exist: the quasi biweekly (12–25 days) time scale (Chen and Chen 1995) and the 30–60-day tropical Madden and Julian (1972, 1994) Oscillation (MJO), which is a climate mode influenced by the outgoing long-wave radiation (OLR) and the atmospheric circulation in the tropics (Wang et al. 2009). On the inter-annual time scale, the variability of monsoon intensity is influenced by ENSO events and the position of the Inter-Tropical Convergence Zone (Dippner et al. 2013). On inter-decadal scales, the variability of circulation is forced by the PDO (Qu et al. 2005), by the Tropical Pacific Decadal Variability (TPDV; Tourre et al. 2001) and by the variability of the WTP warm pool.

The Pacific North Equatorial Current (NEC), located between the subtropical and tropical gyre, bifurcates west of
the Philippines into the northward Kuroshio and the southward Mindanao Current (Fig. 1; Qu et al. 2009; Tozuka et al. 2009; Yan et al. 2013). The latitude of NEC bifurcation is mainly related to ENSO on an inter-annual time scale (Zhai et al. 2014), whereas the decadal variation is highly positively correlated with the TPDV (Tourre et al. 2001). The bifurcation occurs at ~15.5°N on annual average with a band width of fluctuation in bifurcation latitude from 13°N near the surface to 20°N at 800-m water depth (Kim et al. 2004), whereas Hu et al. (2015) gave a band width in bifurcation latitude from 13°N to 17°N. The depth of the subsurface salinity maximum (120–150 m) in the WTP is shallower in the northern basin and deeper in the southern basin (Zeng et al. 2016). During El Niño years, the NEC bifurcation generally occurs at northern latitudes (Kim et al. 2004), resulting in a weaker Kuroshio and a favorable inflow condition of NPEW (Chu and Li 2000; Qu et al. 2005; Zhai et al. 2014).

Based on satellite data and in situ observations, the LST shows a pronounced decadal variation in subsurface salinity (Zeng et al. 2016) with periods of freshening between 1990 and 2012 (Zeng et al. 2014; Nan et al. 2016) and an ongoing period of salinification since 2012 (Zeng et al. 2018). However, no consensus exists on the mechanisms behind this phenomenon. Various authors addressed LST variability...
to ENSO (Zhai et al. 2014; Qu et al. 2004) whereas others pointed to PDO (Zeng et al. 2018; 2016; 2014). Yet another group applied the Island Rule (Godfrey 1989) and concluded that wind stress over the whole equatorial Pacific is the key factor regulating the inter-annual variability of the LST (Wang et al. 2006; Yu and Qu 2013).

Both climate modes, ENSO and PDO, have a similar impact on the Pacific Ocean (Mantua 2001); therefore, separating the dynamic when both modes are in the same phase is difficult. Decadal variations in the sea surface salinity, precipitation, and horizontal and vertical advection during El Niño (La Niña) events show remarkable similarities with those occurring during the positive (negative) phases of the PDO (Delcroix et al. 2007). Evidence also exists that low-frequency climate fluctuation of PDO modulates El Niño intensity (Gershunov and Barnett 1998; Barnett et al. 1999; Torrence and Webster 1999; Tourre et al. 2001; 2005). In addition, it is also not trivial to separate the dynamic of PDO and TPDV because different lead-lag correlations indicated high correlations between PDO and TPDV (Zhai et al. 2014).

The SCS is located between the four monsoon subsystems: the subtropical East Asian monsoon (EAM), the tropical Indian monsoon, the tropical western North Pacific monsoon, and the Australian monsoon (Wang et al. 2009). Unfortunately, more than 20 definitions exist for EAM (Wang and Fan 1999; Chen et al. 2000, 2013; Huang et al. 2003; Wang et al. 2004 and references therein). Wang et al. (2004) pointed out that all definitions vary with respect to the considered variables, including pressure levels, seasons, periods, areas, and data sets. The consequence is that no consistent definition of EAM exists, which may contribute to the variety on LST variability.

This study asks the following scientific questions: what drives this increase in MSW salinity and is it a transient phenomenon? To address these questions, we compute lag correlations between the MSW salinity and different climate modes, and we present surface geostrophic current products from satellite for selected periods (QUID 2018). We additionally use National Center of Environmental Prediction/National Center of Atmospheric Research (NCEP/NCAR) reanalysis data (Kalnay et al. 1996) to construct various composites and an EAM winter index (EAWMI) to analyze the potential reasons for the increase in MSW salinity and the role of different climate time series in influencing its inter-annual variability.

### 2 Material and methods

Six cruises were conducted on the Vietnamese shelf in the SCS (8°–14°N and 105°–111°E) in an effort to characterize the local water masses and seasonal upwelling region (Fig. 2). Five of the six cruises took place between July 2003 and April 2006 aboard the “MV Nghien Cuu Bien” and “RV Sonne,” whereas the sixth occurred in June 2016 aboard the “RV Falkor.” In total, the expeditions covered a range of different seasonal monsoon conditions, as well as post-El Niño climatic conditions in 2003 and 2016 (Table 1). Sea-Bird Scientific CTD systems were used on all cruises with calibrated sensors. Data were validated using a salinometer. The TS-diagram and the mean MSW salinity of all casts and cruises are shown in Fig. 3 and Fig. S1 in the Supplementary Information (SI).

Monthly mean values of the PDO index, the Niño3.4 index, the Multivariate ENSO Index (MEI), the TPDV index, and the El Niño Modoki index (EMI) were used to determine the influence of the WTP variability on LST. The PDO index (Hare and Mantua 2000; Mantua and Hare 2002) is derived as the leading empirical orthogonal function (EOF) of monthly sea surface temperature anomalies (SSTA) in the North Pacific Ocean, poleward of 20°N. The Niño3.4 index is the area averaged monthly SSTA from 5°S–5°N and 170°–120°W calculated from the HadISST1 dataset (Rayner et al. 2003). The MEI is based on six primary variables observed over the tropical Pacific: sea-level pressure (SLP), zonal and meridional components of the surface wind, SST, surface air temperature, and total cloudiness fraction of the sky. The MEI is calculated as the first unrotated EOF of all six observed fields combined (Wolter 1987). The TPDV index is defined as the time coefficients of the first EOF mode of the decadal SSTA in the tropical Pacific Ocean between 120°–280°E and 25°S–25°N (Zhai et al. 2014). Unfortunately, this time series is in the NCEP/NCAR reanalysis data (Kalnay et al. 1996).

| Ship/Cruise | Cruise date | Season/climatic period | No. of stations | No. of CTDs |
|-------------|-------------|------------------------|----------------|-------------|
| MV Nghien Cuu Bien VG-3 | July 2003 | SWM, post-ENSO | 38 | 74 |
| MV Nghien Cuu Bien VG-4 | April 2004 | Spring IM | 38 | 71 |
| MV Nghien Cuu Bien VG-7 | July 2004 | SWM | 34 | 45 |
| MV Nghien Cuu Bien VG-8 | March 2005 | NEM (decaying) | 22 | 41 |
| RV Sonne SO-187–2 | April 2006 | Spring IM | 68 | 100 |
| RV Falkor FK160603 | June 2016 | SWM, post-ENSO | 23 | 55 |
| **Total** | | | **223** | **386** |
NCAR reanalysis only available until 2008. Therefore, we define here the TPDV index as the mean decadal STTA over the same region. EMI as an index of intensity of the central Pacific El Niño is computed using the tripolar procedure of Ashok et al. (2007): 

\[ \text{EMI} = \frac{1}{3} \left[ \text{SSTA}_A - 0.5 \times \text{SSTA}_B - 0.5 \times \text{SSTA}_C \right] \]

The brackets represent the area-averaged SSTA in the region A (165°E–140°W, 10°S–10°N), B (110°–70°W, 15°S–5°N), and C (125°–145°E, 10°S–20°N). These climate modes on a monthly base and their lag correlation to the monthly MSW salinity are displayed in Fig. 4.

NCEP/NCAR reanalysis data (Kalnay et al. 1996) were used to construct an EAM winter index (EAWMI) and composites of geo-potential height (GPH) anomalies and vector wind anomalies at the 850 hPa level for SWM (May to September) and NEM (November to February). We analyzed winter as well as summer composites because the LST has the strongest signals during these periods (Qu et al. 2000).

We selected the area 0°–40°N and 100°–160°E for the following reasons (Fig. 1): it encompasses the area affected by the EAM, including the SCS (10°–25°N, 110°–130°E) and the East China Sea (25°–40°N, 120°–140°N; Chen et al. 2000); it covers the area of the western North Pacific monsoon (5°–22.5°N, 105°–150°E; Wang et al. 2009), the western Pacific subtropical high, and the northern hemisphere part of the WTP warm pool (Huang et al. 2003) and its split region between NEC and NECC (Hu et al. 2017). The area also covers the so-called outcrop zone (25°–35°N, 130°–160°E) where a subsurface salinity anomaly propagates southwestward to the east of Luzon Strait (Yan et al. 2013). The eastward extension of the area is 160°E, which is approximately the position of the nodal line between the area of positive and negative anomalies of the Southern Oscillation (Bjerknes 1969).

Following Chen et al. (2013), we constructed the winter (December, January, and February; DJF) monsoon index as follows. We used NCEP/NCAR reanalysis data (Kalnay et al. 1996) of the meridional wind speed anomalies at 850 hPa level averaged over the areas influenced by the EAM, the SCS, and the East China Sea. The meridional component was used because it has a higher variability than the zonal wind component (Huang et al. 2003). Similar to Chen et al. (2013), we separated the EAWMI time series into ENSO-related and ENSO-unrelated parts. The ENSO-related part was calculated by a linear regression of the time series with respect to the winter MEI. The ENSO-unrelated part is the residuum between the total index and its ENSO-related part, which was then correlated to the winter PDO (Fig. 5).

The composites of GPH and vector winds at the 850 hPa level (Fig. 6 and Figs. S2–S9) in SI were constructed for years with extreme PDO and ENSO values. PDO + values occurred in 1959, 1967, 1986, 1995, and 2004 and PDO − values in 1955, 1963, 1972, 1991, and 2000 (Chan and Zhou 2005; Chen et al. 2013; Yu and Qu 2013). ENSO + values occurred in 1973, 1983, 1992, 1998, 2010, and 2016, whereas ENSO − values occurred in 1956, 1974, 1976, 1989, 2000, and 2011. Furthermore, we constructed composites with respect to the polarity of PDO. Table 2 displays the considered years. According to Gershunov and Barnett (1998), extreme ENSO events were conservatively defined to occur when DJF mean values of both the Niño3.4 and MEI indices deviate more than 0.8 standard deviations from the long-term mean. The ENSO + and ENSO − events were related to the positive and negative phases of PDO. During the period from 1948 to 2017, different PDO epochs were identified: 1947–1976 was a PDO + phase, followed by...
a 1977–1993 PDO + phase (Gershunov and Barnett 1998), a 1993–2012 PDO − phase (Nan et al. 2016; Zeng et al. 2016), and a PDO + phase from 2012 up to today (Zeng et al. 2018). These classifications of the PDO polarity are a result of decadal filtering and therefore not visible in our monthly data (Fig. 4a).

To understand the complex dynamics of WTP and the difference between the post-ENSO + years 2003 and 2016, we investigated surface velocities, MJO time series, and spatial structure of OLR anomalies and 850 hPa vector wind anomalies for the specific years. Daily gridded surface geostrophic velocities were obtained from the EU Copernicus Marine Service Global Ocean Gridded L4 Sea Surface Heights Product (QUID 2018). This product is produced by the SL-TAC multi-mission altimeter data processing system using data from all altimeter missions: Jason-3, Sentinel-3A, HY-2A, Saral/AltiKa, Cryosat-2, Jason-2, Jason-1, T/P, ENVISAT, GFO, ERS1/2. The geostrophic current products were computed using a 9-point stencil width method (Arbic et al. 2012) for latitudes outside of a ±5° band. In the equatorial band, the Lagerloef method introducing the β-plane approximation was used (Lagerloef et al. 1999). Details on the procedure, including how the products were processed, are given in Pujol et al. (2016). The daily velocity fields, which have a spatial resolution of 0.25°, were extracted and averaged for winter (DJF) monsoon and the onset of summer monsoon (MJJ) in early summer for 2003 and 2016 (Fig. 7).

In addition, to understand the difference between 2003 and 2016, we construct fields for winter (DJF) and early summer (MJJ) anomalies of surface OLR (Fig. 8) and vector wind at 850 hPa (Fig. 9) using NCEP/NCAR reanalysis data (Kalnay et al. 1996).
3 Results

The TS-diagram (Fig. 3, Fig. S1 in SI) shows a much higher salinity in waters encountered during the 2016 “RV Falkor” cruise compared to those in the previous cruises. This increase is especially pronounced in the mean salinity of MSW (Fig. 3), which is the core water of the Vietnamese coastal upwelling. The broadening of the MSW salinity range in the TS-diagram between the 24 and 25 isopycnals of $\sigma_t$ densities represents variability in the water mass, which can be attributed to the mechanism of differential mixing of its source waters (Siedler 1970), NPEW and WNPCW. Observations (Tomczak and Godfrey 1994) indicate that NPEW has a core salinity $S > 34.55$ and for the
Fig. 6 Composites of GPH anomalies and vector winds at the 850 hPa level during southeast monsoon for ENSO+ and different PDO polarity (ENSO + PDO +) and (ENSO + PDO −). The contour interval is 1 m
same water mass, Qu et al. (1999) gave a band width of $34.75 < S < 35.25$. In contrast, WNPCW has a core salinity of $S < 34.4$ (Tomczak and Godfrey 1994). The absolute values of mean MSW salinity are thus remarkable: in 2016, the mean value was 34.63, allowing for the conclusion that the LST consisted only of NPEW. The mean MSW salinity was $\sim 34.5$ between 2003 and 2006, which might indicate a specific amount of mixing between WNPCW and NPEW east of Luzon. This is supported by the salinity budget of Zeng et al. (2016) who suggested that entrainment from the mixed layer played a more important role in the freshening period than in the salinifying period.

The LST propagates from the Luzon Strait along the shelf edge of the SCS $\sim 1970$ km to reach the center of the upwelling at $12^\circ$N, where its signal can be detected in MSW. This means that after passing the Luzon Strait, the LST and later the SCSTF reaches the Vietnamese upwelling area 2 to 3 months later, which requires mean travelling speeds of $\sim 0.38$ m/s and $\sim 0.25$ m/s, respectively. Such speeds are reasonable given that observations in the upwelling area showed a meridional southward transport of up to 1.4 m/s at $12^\circ 40'N$ at a water depth of 80–120 m (Dippner et al. 2007), which covers the MSW located in a water depth of 70–140 m.

Five different time series, the MEI, the Niño3.4 index, the PDO index, the TPDV index, and the EMI (Fig. 4), were lag correlated with the monthly mean MSW salinity. Only a weak correlation exists between PDO index and monthly mean MSW salinity. EMI and mean MSW salinity are uncorrelated. The correlation between mean MSW salinity and monthly MEI values was highest with a lag of two months ($r = 0.99$, $p < 0.001$, $N = 6$), though a lag of 3 months was also significant with respect to the 99% confidence level (Fig. 4). A significant longer lasting correlation between MEI, Niño3.4 index, TPDV index, and mean MSW salinity can be identified, which persists nearly half a year. This finding is consistent with the results of Qu et al. (2004) and of Gordon et al. (2012) who documented that LST is the key process conveying the impact of ENSO.
to the SCS. In other words, the inflow into SCS is mainly driven by the variability in the north-western tropic Pacific Ocean.

The EAWMI, the ENSO-related part of EAWMI, and the residuum are displayed in Fig. 5. EAWMI is significantly correlated to the winter MEI \( (r = 0.68, \ p < 0.01, \ N = 68) \). The explained variance of the ENSO-related part and the ENSO-unrelated part is 53% and 47%, respectively. The ENSO-unrelated part is significantly correlated to the winter PDO \( (r = -0.36, \ p < 0.05, \ N = 68) \), which suggests that both climate modes contribute equally to the variability of EAM during winter.

The composites (Figs. S2 and S3) show weak gradients in GPH anomalies for the PDO+ and PDO− pattern indicating no pronounced contribution of PDO to the atmospheric circulation during either the winter or summer monsoons.

If ENSO is included in the SWM composites, all patterns are very similar in the ENSO+ case (Fig. 6). The position of the western Pacific tropical high is nearly the same, but the gradients are different. The highest gradients in GPH

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**Fig. 8** Mean OLR anomalies for winter (DJF) 2003 and 2016 (upper panel) and for the SW monsoon onset in early summer (MJJ) 2003 and 2016 (lower panel)
anomalies occur in the (ENSO + PDO−) pattern, a situation similar to 2003, followed by the ENSO+ pattern. The weakest gradient appears in the (ENSO + PDO +) pattern when ENSO and PDO are in the same phase, e.g., in 2016. The consequence of the similarity in patterns is that the atmospheric forcing of all three patterns causes a northward shift of NEC and the position of the NEC bifurcation latitude. Except the position of the western Pacific tropical high, the pattern during winter monsoon is very similar (Figs. S4 and S6).

The ENSO− and the (ENSO− PDO−) patterns (Figs. S5 and S9) are similar to one other and are characterized by a western Pacific subtropical high located between 30°N–40°N and 130°E–150°E, which overlaps with the

Fig. 9 Mean vector wind anomalies at 850 hPa for winter (DJF) 2003 and 2016 (upper panel) and for the SW monsoon onset in early summer (MJJ) 2003 and 2016 (lower panel)
location of the outcrop zone (Fig. 1; Yan et al. 2013). This atmospheric circulation pattern causes a southward shift of NEC and NEC bifurcation latitude, which drives the Subtropical Counter Current (STCC) at the surface from the outcrop zone into the area of the Luzon Strait.

The (ENSO– PDO +) patterns are remarkable (Figs. S7 and S9). During summer, a large persistent trough south of China prevents inflow into the SCS. During winter, the (ENSO– PDO+) pattern is a particularly striking composite as it shows that a large trough is formed west of the Philippines. This atmospheric forcing causes transport of WNPCW in southwest direction to the Luzon Strait. However, this pattern occurred only five times during the last century (Gershunov and Barnett 1998).

The geostrophic surface velocities (Fig. 7 and Figs. S10–S12 in SI) generally show similar patterns to one another with few specific but essential differences. In all figures, a meandering strong Kuroshio is visible and during winter time, a strong NECC. During early summer, NECC is pronounced in 2003 but weak in 2016. NEC and the NEC bifurcation show different positions. During winter 2003, NEC is located between 15°N and 16.2°N, in winter 2016 between 16.5°N and 17.8°N, in early summer 2003 around 15°N, and in early summer 2016 between 12.5°N and 13.8°N. This indicates that after the SWM onset in 2016, the Kuroshio transports NPEW to the north. In all patterns except early summer 2016, the Kuroshio water is mixed eastward of the Strait of Luzon with WNPCW, which is transported by STCC. In all cases, a mixture of NPEW and WNPCW is the source of LST and hence MSW, but in early summer 2016, only NPEW is transported into the SCS.

A further exciting aspect of our observations is that both 2003 and 2016 were post-ENSO + years, but with different PDO polarity: a negative PDO polarity in 2003 (Nan et al. 2016) and a positive PDO polarity in 2016 (Zeng et al. 2016). Furthermore, in 2003, an El Niño Modoki occurred in the central Pacific (Lee and McPhaden 2010) whereas in 2016, the canonical Eastern-Pacific El Niño occurred (Fig. S13).

Further comparison of 2003 and 2016 shows that the dominant atmospheric driver is the variability in outgoing long-wave radiation (OLR) and in zonal wind in the tropics. During winter in both years, the OLR anomalies were positive but ~35 W/m² higher in 2016 (Fig. 8). The negative anomaly of OLR during early summer 2003 south of 15°N in the WTP indicates strong convection. During early summer 2016, the OLR anomalies were positive in the WTP.

The corresponding 850 hPa vector wind anomalies (Fig. 9) show patchy patterns especially during early summer. In 2003, west of the Philippines and south of 15°N, strong westerly vector wind anomalies at 850 hPa occur as a response to the strong convection. The easterly wind anomaly north of 15°N causes a rather heterogeneous flow field in the geostrophic surface velocity. In contrast, the 850 hPa vector wind anomalies in early summer 2016 show a pronounced easterly component reaching from the equator at 160°E to 20°N close to the Strait of Luzon. This local wind field drives the NEC and the Kuroshio and transports NPEW to the north, feeding the LST.

4 Discussion

The dominant modes of inter-decadal and inter-annual variability in the North Pacific Ocean are the PDO (Mantua 2001) and the ENSO (Philander 1990), which influence the SST, SLP, and surface winds in a similar way. ENSO is the most energetic climate signal in the tropical Pacific with period of approximately 3–5 years (Philander 1990). In contrast, PDO can persist 20–30 years (Latif and Barnett 1994; Mantua 2001; Tourre et al. 2005). Mantua et al. (1997) also pointed out that the response of Pacific climate to ENSO and PDO is stronger in the tropics and weaker in the mid-latitudes.

A decomposition of the constructed EAWMI into its ENSO-related and ENSO-unrelated parts suggests that EAWMI is driven by both ENSO and PDO. The EAWMI is similar to the index constructed by Huang et al. (2003) who showed that the EAM was stronger from the mid-1970s to the late 1980s, but tended to be weaker by the early 1990s, which is consistent with a negative PDO polarity starting in 1993 (Nan et al. 2016; Zeng et al. 2016) and a period of freshening of the LST in the period between 1990 and 2012 (Zeng et al. 2016). The EAWMI is also similar to the winter index constructed by Chen et al. (2000), who showed on a 500 hPa level that EAWMI peaks synchronously with the El Niño events. They also showed that after a strong EAM during winter, the Pacific subtropical high will shift northward in the following summer. In contrast, our results are different to the analysis of Chen et al. (2013), who showed a dominance of the ENSO-unrelated part with an explained variance of 65% and a smaller contribution of the ENSO-related part of 35%. The differences in the two analyses are the length of the time series and the size of the considered area, which suggests that the definition of EAWMI and its ENSO relation is very sensitive to the selection of the considered area. Chen et al. (2013) used an area that covers 20°N–40°N in meridional direction and excludes a major part of the tropics. It might be that this affects the dominance of the ENSO-unrelated part in the EAWMI.

On a seasonal scale, the EAM and in particular the position of the Inter-Tropical Convergence Zone are the major driving forces of circulation and hydrography in the SCS (Dippner et al. 2013; Chen et al. 2013). One of the strongest response signals is the position and intensity of coastal upwelling off the coast of Vietnam, which in turn influences the pathway of the Mekong River plume.
of the Indo-Pacific warm pool, the largest area of warm surface water in the world ocean. This region is characterized by a SST exceeding 28 °C, weak trade winds, and atmospheric deep convection (Kidwell et al. 2017; Weller et al. 2016). WTP can be classified by different properties such as centroid movement, length, depth, heat content, and volume (Kidwell et al. 2017), which are correlated to all equatorial climate indices and the PDO.

The influence of zonal winds and OLR anomalies close to the equator, the essential variables for the MJO, contributes to more than 55% of the atmospheric anomalies over the tropics (Kessler 2001; Roxy et al. 2019). This aspect of potential forcing on the LST intrusion dynamics has not been considered as far as we know. In 2003, a strong convection (negative OLR anomalies, Fig. 8) occurred during the SWM onset in early summer. When the convection is strong, the convergence of flow is intensified in the lower troposphere and a cyclonic circulation will be intensified over the WTP (Fig. 9; Huang et al. 2001). In 2016, the convection in early summer was weak over the WTP (Fig. 8). This will in turn intensify an anticyclonic circulation over the WTP (Fig. 9), which transports the NPEW into the SCS (Fig. 7). These results are also consistent with the MJO variability in the WTP (Figs. S14 and S15 in SI) as seen in the MJO index of Wheeler and Hendon (2004) for these years. Summing up these findings, we can conclude that the increase in the salinity of MSW in the SCS in 2016 was a result of multi-time scale processes. The inter-annual variability of LST is dynamically related to the Kuroshio and the NEC bifurcation latitude (Wang et al. 2006), which is regulated by ENSO. When ENSO and PDO are in a positive phase, the Kuroshio is much weaker after the NEC bifurcation. The consequence is that the meridional advection of potential vorticity is not strong enough to overcome the β-effect, allowing the Kuroshio to transport more NPEW through the Luzon Strait, the so-called teapot effect (Sheremet 2001). Model simulations indicate that the LST would be greatly reduced without the β-effect (Yuan 2002).

The ENSO+ and the (ENSO+PDO+) patterns as well as the ENSO− and the (ENSO−PDO−) patterns are very similar during winter and summer monsoon (Figs. S4–S9). This is in good agreement with the results of Mantua (2001) and Delcroix et al. (2007) who demonstrated that both climate modes show remarkable similarities if both modes are in the same positive or negative phase.

All ENSO+ composites independent of the PDO phase shift the bifurcation latitude to the north, which is a necessary condition of LST intrusion into the SCS. However, these composites do not explain the difference between 2003 and 2016, two post-ENSO+ years with different PDO polarity and why during the 2016 SWM mainly NPEW was transported into the SCS (Fig. 7).

To understand the mechanisms, we investigated the complex multi-time scale processes in the WTP, which is a part of the Indo-Pacific warm pool, the largest area of warm surface water in the world ocean. This region is characterized by a SST exceeding 28 °C, weak trade winds, and atmospheric deep convection (Kidwell et al. 2017; Weller et al. 2016).
excluded from the calculation. Considering the narrowness and the shallowness of the Strait of Luzon, friction cannot be neglected. The application of Island Rule (Wang et al. 2006; Yu and Qu 2013) might yield qualitative estimates of LST but cannot explain the transport of NPEW into the SCS.

Is the increase in MSW salinity a transient phenomenon? To investigate this aspect, we searched for periods with conditions similar to those in 2016. An inspection of the NCEP/NCAR reanalysis data (Kalnay et al. 1996) showed that only one record exists that fulfills the criteria of a post-ENSO + year with positive PDO phase and weak MJO during the SWM onset. The year 1983 is characterized by positive anomalies in OLR and easterly winds in the WTP (not shown), which were similar to the atmospheric conditions in 2016. However, we cannot show that in this year, MSW has a higher salinity than normal. The reason is that no observations are available because in situ observations in SCS were sparse and infrequent before the Vietnamese-German cruises, which started in 2003. In addition, we cannot show that during the 1983 SWM NEC transported NPEW into the SCS, because no geostrophic velocities from satellite were available since the observation program started in 1993 (QUID 2018). We can identify similar atmospheric conditions in the past as in 2016, but we cannot show the response of the WTP and SCS to this forcing. The consequence is that we cannot answer questions on the transience of this phenomenon.

Is the increase in MSW salinity related to global warming? During the last decades, various observations indicated major changes in the Pacific Ocean especially in the WTP. From 1980 to 2010, the increasing intensity and the occurrence frequency of El Niño Modoki events in the central Pacific almost doubled (Lee and McPhaden 2010). Model inter-comparison indicated that the occurrence ratio of El Niño Modoki to the canonical El Niño is projected to increase as much as five times under global warming (Yeh et al. 2009). The change is related to a flattening of the thermocline in the equatorial Pacific. Weller et al. (2016) address the observed increase in intensity and frequency to greenhouse gas forcing. Observed Indo-Pacific sea level pressure reveals a weakening of the Walker circulation, which has altered the thermal structure and circulation of the tropical Pacific (Vecchi et al. 2006). Model simulations indicate that this trend is related to anthropogenic forcing (Vecchi et al. 2006). A reconstruction of MJO over 1905–2008 using tropical surface pressure showed a 13% increase per century in MJO amplitude (Oliver and Thompson, 2012). For the period 1981–2018, the rapid warming over the tropical oceans has warped the MJO life cycle with increasing residence time over the Indo-Pacific maritime continent of 5–6 days. These changes in MJO life cycle were associated with an expansion to the Indo-Pacific warm pool (Roxy et al. 2019). All of these changes in the last decades might indicate the consequences of global warming. The increase in the salinity of MSW in the SCS may be a small overseen piece of puzzle in this context and should be the subject of further observations.

The MSW sampled in summer 2016 had a significantly higher salinity compared to previous years, which resulted from a particularly strong El Niño during the winter of 2015/2016. This salinity shift in MSW necessitates a modification of the water mass definition given by Dipperner and Loick–Wilde (2011); the changes to which are displayed in Table 3. Note that WM1 to WM4 are not characteristic water masses in the classical definition of Helland–Hansen (1916); however, an introduction of these mixed water masses is meaningful for various reasons. Firstly, the temperature and salinity ranges of the water masses defined by Rojana–anawat et al. (2001) are too coarse to define “end members” of mixing. Secondly, although characteristic water mass analysis is a useful tool in understanding mixing dynamics of water masses (Wüst 1936), the mixing between MSW and OSW is difficult to interpret because of a clear bifurcation in the mixing diagram (Dipperner et al. 2007). Thirdly, all water masses are different in their biogeochemical and phytoplankton species compositions (Loick–Wilde et al. 2017), which justify the introduction of mixed water masses or alternatively the habitat type concept (Weber et al. 2019). Our modified definition of characteristic water masses may be useful for oceanographers and biologists operating in the SCS.

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