A reduction in the sea surface warming rate in the South China Sea during 1999–2010

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Abstract

The South China Sea (SCS) experienced a significant reduction in warming rate ($-0.01\text{ }^{\circ}\text{C decade}^{-1}$, $p > 0.10$) during 1999–2010 following an accelerated and unprecedented warming ($+0.15\text{ }^{\circ}\text{C decade}^{-1}$, $p < 0.01$) in the last three decades (1970–1998). However, most global climate models of the CMIP5 RCP4.5 scenario failed to capture this SCS warming slowdown. In this study, we identify two drivers through numerical simulations by using a regional high-resolution, ocean–atmosphere coupled climate model: the major variance (75%) in the sea surface warming slowdown could be explained by the strengthened winter monsoon over the SCS, and the minor variance (12%) could be explained by the changes in the upper ocean circulations. The winter monsoon over the SCS is likely linked to the La Niña-like SST pattern in the eastern tropical Pacific, which strengthens the Walker circulation and results in anticyclonic circulation over the northwestern Pacific. This enhanced winter monsoon is the atmospheric driver that slows down the SCS basin-scale warming, while the largest reduction of the warming rate occurs in the northern SCS that can be attributed to the oceanic throughflow via the Luzon Strait. These findings could have important implications for future climate projections over the SCS and adjacent oceans.

1 Introduction

The South China Sea (SCS), surrounded by 13 Southeast Asian countries that are home to over 600 million people, is strategically crucial in shipping goods, lucrative fisheries-based food security, and large oil–gas reserves. Hence, the spatiotemporal variability in the SCS ocean–climate system has immeasurable impacts on human economic activities (He et al. 2014; Tong et al. 2017), ecosystem stability,

natural disasters, such as typhoon intensity (Guo and Tan 2018), and SCS summer monsoon onset (Wu 2010).

Embedded between the tropical Indian and Pacific Oceans, the complex climate variability in the SCS is driven by both local air–sea interactions and large-scale remote forcing through oceanic and atmospheric bridges (Gordon 1986; Klein et al. 1999; Wei et al. 2016a). The SCS is injected with western tropical Pacific water by the Luzon Strait and exports water through the Karimata Strait into the Java Sea (Du and Qu 2010; Qu 2000) and the Mindoro–Sibutu pathway into the Sulu Sea (Gordon et al. 2012; Jiang et al. 2019; Li et al. 2018, 2019, 2020; Metzger and Hurlburt 1996). The SCS experiences strong seasonality due to variability in solar radiation and monsoonal forcing (Chen et al. 2003b; He and Wu 2013; Liu et al. 2001; Qu 2001); in addition, there is interannual variability induced by remote forcing from tropical oceans, such as the El Niño–Southern Oscillation (ENSO) (Liu and Xie 1999; Wei et al. 2016a). On a longer timescale, the SCS has experienced accelerated and unprecedented warming since the 1970s. However, the SCS warming was believed to continue until recently according to many previous studies (Lin et al. 2011a, b; Thompson et al. 2017; Xiao et al. 2019; Yang and Wu 2012; Wei et al. 2013), as few of them focused on the SCS warming

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slowdown during 1999–2010, and the mechanisms have not been well addressed.

In the two decades following 1999, a so-called “hiatus” or slowdown of global surface warming was found (Abraham et al. 2013; Balmaseda et al. 2013; Levitus et al. 2009; Meehl et al. 2011). One explanation relates the “hiatus” to the Pacific Decadal Oscillation (PDO) and decadal ENSO signal (Lin et al. 2011a, b; Trenberth and Fasullo 2013). These studies found that the PDO shifted from a positive phase during 1976–1998 (warmer Indo-Pacific pool and cooler Western North Pacific) to a negative phase after 1999 (cooler Indo-Pacific pool and warmer Western North Pacific). During the PDO negative phase, the decadal Niña 3.4 index shows a predominant La Niña-like phase, which enhances trade winds and cools the eastern tropical Pacific. Given that the tropical Pacific largely accounts for the global surface temperature variation, the La Niña-like phase could make a significant contribution to the “hiatus” (Kosaka and Xie 2013). Other explanations, such as heat sinking into the deeper ocean (Chen and Tung 2014), natural variability and radiative forcing of the climate system (Dai et al. 2015; Huber and Knutti 2014; Watanabe et al. 2014), volcanic eruption (Santer et al. 2014), and increased Indonesian throughflow (ITF), which transfers Pacific heat into the Indian Ocean (Lee et al. 2015), were also found, arguing that global warming forced by a long-lasting surging trend of anthropogenic aerosol emissions will persist. Some climate model projections suggest that the “hiatus” may continue to the end of the 2020s (England et al. 2014; Medhaug et al. 2017).

The “hiatus” of global surface warming has been shown to have great impacts on regional climate, such as the reversal of Indian monsoons (Jin and Wang 2017), amplified Arctic warming (McGregor et al. 2014), and extreme weather events (Sillmann et al. 2014). In this study, we identified a significant warming slowdown in the SCS during 1999–2010 (Fig. 1a) based on multiple datasets, whereas most of the Coupled Model Intercomparison Project Phase 5 (CMIP5) models failed to capture this regional warming slowdown. With a set of sensitivity experiments using a high-resolution, regional coupled model, the causes of the SCS sea surface warming slowdown during 1999–2010 are analyzed.

2 Data and methods

2.1 Observational and reanalysis datasets

SST data products derived from satellite retrievals, observations, and reanalysis are used in this study. Winds and air temperatures at various vertical atmospheric levels are from National Centers for Environmental Prediction (NCEP) reanalysis products and coupled model intercomparison project

phase 5 (CMIP5) simulations. The SST datasets are collected from the Simple Ocean Data Assimilation ocean/sea ice reanalysis (SODA) v3.4.2 ($0.25^\circ \times 0.25^\circ$, Carton et al. 2018) and v2.2.4 ($0.4^\circ \times 0.4^\circ$, Carton and Giese 2008), the Hadley Centre Sea Ice and SST data set (HadISST; $1^\circ \times 1^\circ$, Rayner et al. 2003), and the Extended Reconstructed SST (ERSST) v5 data set ($2^\circ \times 2^\circ$, Smith et al. 2008). SODA is ocean reanalysis data, including various observational datasets with data assimilation methods. HadISST is a global telecommunications system (GTS)-based dataset. ERSST is a global monthly SST analysis from the International Comprehensive Ocean–Atmosphere Dataset, which does not include satellite data. The air temperatures and winds at the 200 hPa and 850 hPa pressure levels are from the NCEP/NCAR Reanalysis 1 ($2.5^\circ \times 2.5^\circ$, Kalnay et al. 1996). Simulations of SST from 14 CMIP5 simulations with 53 runs are also included (Table 1). The CMIP5 historical runs cover the period of 1970–2002, while for the 2002–2015 period, some historical runs are merged with the so-called RCP4.5 (a medium mitigation scenario called ‘Representative Concentration Pathway 4.5’) emission scenarios (Taylor et al. 2012). Note that anomaly computation is conducted by removing the SCS basin average, and the decadal trend of ocean and air values is averaged yearly to remove seasonal variability.

2.2 RegCM-FVCOM coupled model

The regional air-sea fully coupled model for the SCS was developed and validated in our previous studies (Wei et al. 2013, 2014; Jiang et al. 2019). The model adopts Regional Climate Model version 3 (RegCM3) as the atmospheric component (Pal et al. 2007), which was originally developed at the National Center for Atmospheric Research (NCAR), and the Finite-Volume Coastal Ocean Model (FVCOM) is adopted as the oceanic component (Chen et al. 2003a), which was developed at the University of Massachusetts-Dartmouth. The atmospheric component adopts the dynamical core based on the hydrostatic version of Pennsylvania State University/NCAR Mesoscale Model Version 5 (MM5) (Grell et al. 1994) and the atmospheric radiative transfer scheme from NCAR’s Community Climate Model Version 3 (CCM3) (Kiehl et al. 1996). A bulk aerodynamic ocean flux parameterization scheme is used to compute the ocean surface fluxes (Zeng et al. 1998). The oceanic component, with an unstructured (triangular) grid, solves the momentum and thermodynamic equations with a second-order finite-volume flux scheme, which amalgamates the advantages of finite-element methods for geometric flexibility and finite-difference methods for computational efficiency. Vertical eddy viscosity and diffusivity are resolved by the Mellor and Yamada level 2.5 turbulent closure scheme (Mellor and

Fig. 1 Sea surface temperature (SST) time series and spatial trends. **a** Observed and modeled time series of the South China Sea (SCS) SST anomalies ($^{\circ}\text{C}$). The SST is averaged within an area of 2°N – 24°N , 100°E – 120°E , as shown in **b**–**d**. The SODA re-analysis (black), Hadley observed SST (blue), ERSST (green) are served as reference datasets comparing with RegCM3-FVCOM coupled model (red). The linear decadal trends (denoted by ‘rc’) and p values are listed in the box located at the upper-left corner in panel **a**. The dashed black line indicates the trend of SODA. The vertical shading in gray emphasizes the abnormally high SST year of 1998, which was masked out when performing the trends. **b**, **c** Spatial distributions of decadal trends calculated using SODA SST ($^{\circ}\text{C decade}^{-1}$) during 1970–1997 and after 1999, respectively. **d**, **e** Similar to **b**, **c**, but using modeled SST. The black and white dots show the significant trends at 90% significant level

Yamada 1982), and horizontal diffusivity is calculated by the Smagorinsky turbulence closure (Smagorinsky 1963).

The model domains of RegCM3 are set from 85°E to 142°E and from 20°S to 30°N , which is slightly larger than the FVCOM. This model covers the entire SCS basin, East Asian monsoon area, SCS throughflow (SCSTF) and Indonesian throughflow (ITF). The horizontal resolution of RegCM3 is homogeneously 60 km; however, the resolution of the FVCOM varies from ~ 5 km on the shelves, continental slopes and narrow straits to ~ 50 km within the deep SCS and ~ 200 km along the ocean open boundaries. The ERA-40 dataset from the European Centre for Medium-Range Weather Forecasts (ECMWF) (Uppala

et al. 2005) offers the initial and boundary conditions for RegCM3, and the SODA reanalysis product provides the FVCOM model inputs. After a 10-year spin-up during the 1960s, RegCM3 and FVCOM were simultaneously integrated forward from 1970 to 2010 through an OASIS3 coupler, which allowed flexible interpolation and transfer of the coupling fields (winds, heat fluxes and SST) between different grids every 3 h. More detailed model setup and validations can be found in our previous study (Wei et al. 2013). Moreover, the FVCOM-only model is validated against in situ datasets (Xu et al. 2013), and the local air-sea feedback mechanisms with a fully coupled version are also verified in Xue et al. (2014).

2.3 Partial-coupling experiments

To identify the driving factors, the decomposition method with partial-coupling (P–C) experiments was used, following our previous study (Wei et al. 2016a). P–C experiments allow for partial air–sea coupling by controlling either atmospheric or oceanic variables and are widely used in global GCMs (Ding et al. 2014a; Wu et al. 2003; Zhong et al. 2008). The difference between the control run (CTL) and P–C experiments decomposes the effect of the controlled variables. To separate the different roles induced by atmospheric effects and boundary intrusion effects on the SCS, a set of experiments are carried out. The fully coupled model integrated from 1970 to 2010 is referred to as the control run (CTL).

In the temperature tendency equation (Eq. 1),

$$\frac{\partial T}{\partial t} = -\left(u\frac{\partial T}{\partial x} + v\frac{\partial T}{\partial y} + w\frac{\partial T}{\partial z}\right) + \frac{\partial}{\partial x}\left(A_h\frac{\partial T}{\partial x}\right) + \frac{\partial}{\partial y}\left(A_h\frac{\partial T}{\partial y}\right) + \frac{\partial}{\partial z}\left(K_h\frac{\partial T}{\partial z}\right) + \frac{Q}{h\rho_0 C_p}, \quad (1)$$

where the left-hand side is the temperature change rate; u , v and w are the ocean current velocities; A_h and K_h are the horizontal and vertical diffusion coefficients; and h , C_p , and ρ_0 are the mixed-layer depth, water density and specific heat of water, respectively. Therefore, the temperature change is controlled by advection, horizontal and vertical diffusion, and heat flux terms. However, advection and heat flux terms always dominate near the surface mixing layer (Jiang et al. 2018; Uhlhorn and Shay 2013; Wei et al. 2016a), occupying up to 96% of the total change over the whole SCS (see details in Wei et al. 2016a), but the vertical mixing term can be significant at the subsurface, e.g., the thermocline, where the vertical temperature gradient is much larger. Regional climate processes mostly involve boundary-control issues, as are the experiments designed in this study. To separate the effects of atmospheric and oceanic forcing on the SCS SST variance, in the “fixed surface condition” (FSC) experiment, the lateral conditions of the air model (heat fluxes and winds) for 1980–2010 are prescribed repeatedly using the lateral conditions of 1970–1980. Similarly, in the “fixed boundary condition” (FBC) experiment, the lateral boundary conditions of the ocean model (temperature, salinity and transport) are prescribed repeatedly using the boundary conditions of 1970–1980. Apart from the lateral conditions, other settings are kept the same as in the CTL. Thus, the difference between CTL and FSC identifies the SST variance due to the atmospheric surface forcing, and CTL–FSC identifies the SST variance due to oceanic advection.

As a result, the forcing of surface atmospheric effects can be estimated by $CTL - FSC$; the effects of boundary

intrusion, mainly dominated by the variability in the Luzon Strait intrusion, can be estimated by $CTL - FBC$. For simplicity, here, we use FSC and FBC to represent $CTL - FSC$ and $CTL - FBC$, respectively.

3 The remarkable warming slowdown in the SCS

3.1 Warming slowdown in observations

Based on the global reanalysis datasets, Fig. 1a shows that the SCS SST experienced an accelerated and unprecedented warming trend ($+0.15\text{ }^{\circ}\text{C decade}^{-1}$, $p < 0.01$) during 1970–1997. A strong warming trend is detected over

the whole SCS area, from coastal margins to the central realm (Fig. 1b). However, this sea surface warming trend encountered a significant slowdown during 1999–2010, with an overall rate of $-0.01\text{ }^{\circ}\text{C decade}^{-1}$ ($p > 0.10$), and continued to flatten even into 2019. The most remarkable warming slowdown occurs in the northern SCS, with a cooling rate of $-0.5\text{ }^{\circ}\text{C decade}^{-1}$ (Fig. 1c, $p < 0.10$).

To demonstrate that the sea surface warming slowdown marks a decadal feature, an 18-year running mean of SST trends is carried out (Fig. 2). The recalculated SST trends clearly demonstrate that the SCS SST continues to rise during the period of 1970–1990. However, the sea surface warming rate drops from above $0.1\text{ }^{\circ}\text{C decade}^{-1}$ to below zero after 1990 and is persistent in the following years in the SODA, HadISST and ERSST data.

Further analysis on a seasonal timescale based on HadISST shows that the warming slowdown of the SCS sea surface after 1999 is much stronger during boreal winter (drops from 0.25 to $-0.08\text{ }^{\circ}\text{C decade}^{-1}$, $p < 0.10$), followed by spring (drops from 0.20 to $-0.03\text{ }^{\circ}\text{C decade}^{-1}$, $p < 0.15$) and summer with a slight slowdown trend, as shown in Fig. 3a. The spatial pattern shows that during 1970–1997 the SCS warming trend is nearly uniform over the entire SCS in the four seasons, while the warm slowdown during 1999–2015 occurs mainly in the northern SCS in the winter and spring seasons. The largest cooling rate along the northern SCS shelf follows the path of the SCS through from the Luzon Strait to the Vietnam coast. Note that this cooling rate along the northern SCS shelf is significantly weakened in the summer and disappeared in the autumn.

Fig. 2 SST trends calculated using an 18-year moving window. The linear trends ($^{\circ}\text{C decade}^{-1}$) are computed based on the SCS SST time series in Fig. 1a. The significant trends at the 90% significant level are marked with 'square' symbols

3.2 Warming slowdown in models

Given the complex topography of the SCS and its multiple narrow passages to the open western Pacific, global climate models of CMIP5 are usually unable to accurately resolve physical processes in this region due to their coarse resolutions, which could be one of the reasons that most CMIP5 models fail to simulate the SCS SST. To include the local effects from topography, coastlines and islands, we adopted a high-resolution RegCM-FVCOM coupled model. The remarkable warming slowdown in the SCS is reproduced well by the model, with a mean absolute error (MAE) of 0.12°C and correlation coefficient of 0.90 with observational products (see red line in Fig. 1a). Specifically, compared with the SODA data, the coupled model captures the warming trend in the central and southern SCS well from 1970 to 1997 (Fig. 1b, d) and the cooling trend in the northern SCS during 1999–2010. (Fig. 1c, e).

4 Identifying the drivers of the SCS warming slowdown

4.1 Partial-coupling experiments

Previous studies have shown that the semi-closed SCS climate system is largely driven by the variability in East Asia monsoons and Luzon Strait intrusions (Gordon et al. 2012; Klein et al. 1999; Wei et al. 2016a). Therefore, a set of partially coupled experiments are carried out to investigate the contribution of monsoon winds and Luzon Strait transport to the warming slowdown in the SCS. The experiment details are shown in Sect. 2.3.

The two P-C experiments were carried out, and the results are shown in Fig. 4. The total variations in the SCS SST can generally be represented by the two components, explaining over 87% of the total variance (Fig. 4a). The decomposed components show that the change in air surface conditions (CTL-FSC) dominated the remarkable warming slowdown of the SCS, accounting for 75% of the total variance ($-0.17^{\circ}\text{C decade}^{-1}$, $p < 0.10$) during 1999–2010. This result indicates that the SCS warming slowdown is largely attributed to the enhanced monsoons. On the other hand, the SST changes induced by ocean boundary conditions (CTL-FBC) account for 12% of the total variance. The residual ($\sim 13\%$ of the total variance) may come from the nonlinear vertical mixing and vertical advection of ocean subsurface, which cannot be linearly decomposed. Note that the ΔSST shown in Fig. 4a, b is calculated as follows:

$$\begin{aligned} \text{Total } \Delta\text{SST} &= \text{CTL}(1980\text{s}, 1990\text{s}, 2000\text{s}) - \text{CTL}(1970\text{s}); \\ \Delta\text{SST due to air surface forcing} &= \text{CTL}(1980 - 2010) - \text{FSC}(1980 - 2010), \\ \Delta\text{SST due to oceanic boundary} &= \text{CTL}(1980 - 2010) - \text{FBC}(1980 - 2010). \end{aligned}$$

4.2 Attribution of SCS warming slowdown

Figure 4b shows that the sea surface warming slowdown over the SCS in the FSC experiment is much stronger in boreal winter ($-0.33^{\circ}\text{C decade}^{-1}$, $p < 0.10$) than in summer ($-0.09^{\circ}\text{C decade}^{-1}$, $p > 0.10$). The variability in the warming slowdown trends in the boreal winter and summer seasons is consistent with the observational products (Fig. 3a). For the direct cause of the warming slowdown in the FSC experiment, the net heat flux absorbed by the ocean experienced a rapid decrease (-10.8 W m^{-2}

Fig. 3 SST time series and spatial trends of each season. **a** SCS SST (°C) time series of boreal winter (purple), spring (blue), summer (orange) and autumn (yellow). The SST data used here comes from HadISST, similar with the results of ERSST, OISST and SODA reanalysis. The vertical shading in gray is similar with that in Fig. 1. **b**

The SCS SST trends (°C decade⁻¹) of each season during 1970–1997 and 1999–2015. **c–j** The spatial pattern of decadal SST trends of each season during 1970–1997 and 1999–2015 respectively. The black dots indicate the 90% significant level

decade⁻¹, $p < 0.10$) during 1999–2010 (Fig. 5c). The majority of the net heat flux decrease originates from the accelerated latent heat uptake from the ocean surface (r

$= -0.94$; $p < 0.01$), from a negative rate of -0.7 W m^{-2} decade⁻¹ ($p < 0.10$) during 1970–1997 to a positive rate of 8.2 W m^{-2} decade⁻¹ during 1999–2010 ($p < 0.10$,

Fig. 4 Time series and spatial trends of modeled SST and surface wind. **a** Time series of SCS SST difference ($^{\circ}\text{C}$, averaged within the same SCS area as Fig. 1a) in simulations of CTL (black line), fix surface condition experiment (FSC, red line) and fix boundary condition (FBC, blue line). The percentages in legends are contribution to CTL calculated based on ' rc_{FSC}/rc_{CTL} ' and ' rc_{FBC}/rc_{CTL} '. **b** Time series of SST difference ($^{\circ}\text{C}$) in FSC experiment during boreal summer (JJA, red solid line) and boreal winter (DJF, red dashed line) using the left y-axes, and surface absolute wind speed (ms^{-1}) abnormally of boreal winter (green solid line) using the right y-axes (larger values mean stronger southward winter monsoon). **c, d** The spatial distribution of FSC decadal SST trends ($^{\circ}\text{C}$ decade $^{-1}$, shading) and decadal surface wind abnormally trends (ms^{-1} decade $^{-1}$, green arrows) of boreal winter during 1980–1997 and after 1999 respectively. **e, f** Similar to **c–d**, but for FBC experiment in boreal winter (the green arrows indicate the current trends on the ocean surface). The black dots indicate the trends at 90% significant level. More details of ΔSST calculation in **a, b** are shown in Sect. 2.3

Fig. 5b). The accelerated evaporation (latent heat uptake) is closely related to the revival of the modeled East Asian winter monsoon (EAWM), from a negative rate of $-0.21 \text{ ms}^{-1} \text{ decade}^{-1}$ ($p < 0.10$) before 1999 to a positive rate of $0.71 \text{ ms}^{-1} \text{ decade}^{-1}$ after 1999 ($p < 0.10$, Fig. 4b). Such a revival of the modeled EAWM is consistent with the National Centers for Environmental Prediction (NCEP) reanalysis products (Fig. 6). Therefore, the warming

slowdown trend in boreal winter is obviously attributed to the revival of the EAWM ($r = -0.49$, $p < 0.01$, Fig. 4b).

The spatial distribution of the SST trend over the SCS in the FSC experiment shows a basin-wide homogenous warming rate of $0.14 \text{ }^{\circ}\text{C} \text{ decade}^{-1}$ ($p < 0.01$, Fig. 4c), which is coincident with the weakened EAWM over the SCS from 1980 to 1997 (Fig. 4b, green line). However, after 1999, the enhanced EAWM resulted in strong sea surface cooling

Fig. 5 Time series of ΔSST and heat flux. **a** Same as Fig. 4a, but for SST abnormally of FSC + FBC experiments versus CTRL. The black line uses the left y-axes while the red line uses the right y-axes. **b, c** Same as Fig. 4b, but for latent heat flux and net heat flux ($W m^{-2}$) respectively. The red dashed line is for boreal summer and green solid line is for boreal winter

over the entire SCS (Fig. 4d). This result confirms that the EAWM shifting before and after 1999 is the major forcing on the SCS SST, accounting for 75% of the SST variance and influencing the entire SCS basin. On the other hand, the effect of oceanic boundary forcing on the SCS SST is small compared to the EAWM effect, but obviously larger than its effect during 1980–1997 (Fig. 4e, f). As demonstrated in many previous studies, the SCS throughflow via the Luzon Strait strengthened after 1999 when the climate system shifts from positive PDO with more El Niño events to negative PDO with more La Niña events (Qiu and Lukas 1996; Kim et al. 2004; Qiu and Chen 2010; Gordon et al. 2014). Our previous study also confirmed the strengthened SCSTF by decomposing the sea surface height (Wei et al. 2016a, b), which drives more cool water from the northern SCS to the central and southern regions.

To reveal the mechanism of the enhanced EAWM, Fig. 7 shows the spatial trends of surface winds and heat fluxes in the SCS, and Fig. 8 indicates the linkage of the SCS domain to the Pacific. The results demonstrate that the SCS SST is positively correlated with the SST over

the eastern tropical Pacific (Fig. 8), implying that the SCS warming slowdown can likely be linked to the eastern tropical Pacific. Based on the NCEP atmospheric analysis data, Niña-like sea surface cooling in the eastern tropical Pacific results in convection and an anticyclonic anomaly in the western Pacific in the lower atmosphere (Fig. 8b). This phenomenon has been documented in previous studies (Ding et al. 2014b; Liu et al. 2011; Wang et al. 2000). The anticyclonic air circulation anomaly further enhances the EAWM over the SCS (Fig. 7e). In contrast, the SCS sea surface experiences nonsignificant cooling during the boreal summer (Fig. 7a–d), with a pattern balanced by the opposite phase of the net heat flux trend. This result can be explained by a negative feedback that the cold (warm) SST forces less (more) evaporation and thus less (more) highly reflective cirrus clouds form, and consequently, more (less) shortwave radiation reaches the ocean surface (Ramanathan and Collins 1991). This feedback weakens as the Northern Hemisphere enters the boreal winter season, in which the SCS SST is mainly controlled by the revived EAWM.

Fig. 6 The same as Fig. 3, but for surface meridional wind speed abnormally from NCEP dataset. The positive trends mean northward monsoon while negative trends mean southward (e.g. negative trends in winter mean stronger winter monsoon)

5 Summary and discussion

A significant slowdown in the SCS warming rate during 1999–2010 is observed in both remote-sensed satellite-derived products and reanalysis datasets (Fig. 1), following

a long-term warming trend from 1970 to 1997. However, most of the IPCC CMIP5 climate models failed to reproduce this warming slowdown in the SCS. Based on a high-resolution regional air-sea coupled model with a set of P-C

Fig. 7 Spatial trends of wind, heat flux and SST in FSC. **a–d** Decadal spatial trends of SST versus surface wind (**a**), latent heat flux uptake (**b**), short wave heat flux absorbed (**c**) and net heat flux absorbed (**d**) during boreal summer, respectively. **e–h** The same as **a–d**, but for boreal winter

Fig. 8 Correlations between SCS SST and Pacific temperature and wind. **a** Correlation coefficients between SCS SST and NCEP air temperature (shading) and wind vectors (purple arrows) at 200 hPa over Pacific Ocean during boreal winter (DJF) of 1970–2015. **b** Similar with **a**, but for the wind vectors at 850 hPa and SODA SST. Had-ISST and ERSST indicate the same pattern and are not shown here.

The Walker Circulation is highlighted with bigger orange and light blue vectors. The dark blue curves with arrows in **b** indicate the circulation trends at 850 hPa during 1999–2015. The black dots indicate the significant correlation with temperature at the 95% level. Those insignificant correlation coefficient with wind vectors are omitted

Fig. 9 Time series and spatial distribution of eastern Tropical Pacific SST. **a** Observational time series of eastern Tropical Pacific SST anomalies ($^{\circ}\text{C}$) during boreal winter from SODA (black), HadISST (blue) and ERSST (green). Note that the SST is averaged within the same area of nino3.4 (5°N to 5°S , 170°W to 120°W). The linear decadal trend of SODA is plotted as black solid line (regression coefficients denoted by ‘ rc ’ below the lines). As a reference, the extension of the warming trend is plotted as black dashed line after 1997. **b** The distribution of the composite DJF-mean SST anomalies of HadISST during 1999–2015 (similar results exist in SODA and ERSST, not shown here). The bigger arrows are the same as those in Fig. 8

experiments, the model results show that this SCS warming slowdown has the largest magnitude in boreal winter and spring, forced by strengthened EAWM winds during 1999–2010 and thus increased latent heat uptake over the SCS (Figs. 4, 5). The strengthened EAWM may be linked to the La Niña-like SST pattern in the eastern tropical Pacific, which was not simulated by those CMIP5 models. The largest reduction in the SCS warming rate can be attributed to the oceanic throughflow by the Luzon Strait, where most of the CMIP5 models are unable to resolve the local topography and coastlines, as well as the Luzon Strait throughflow.

Increasing local resolution has been adopted in regional models to improve model simulations in the SCS and surrounding oceans and seas (Wei et al. 2013; Lin et al. 2011a,

b; Liu et al. 2011; Van Sebille et al. 2014; Tranchant et al. 2016; Li et al. 2019; Jiang et al. 2019; Xue et al., 2014). However, less attention has been given to the linkage of the SCS to large-scale remote forcing through atmospheric bridges. Figure 8 demonstrates that the SCS SST is positively correlated with the SST over the eastern tropical Pacific and positively correlated with the anticyclonic circulation at 850 hPa in the northwestern Pacific during boreal winter. The warming trend of the SCS sea surface during 1970–1997 was captured well by most CMIP5 models when the El Niño-like SST pattern was dominant in the eastern tropical Pacific. However, an obvious warming slowdown occurred over the eastern tropical Pacific during 1999–2015, resulting in cyclonic circulation at 850 hPa in the northwest

Fig. 10 SCS SST time series and spatial trends of CMIP5 simulations. **a** SCS SST Time series of CMIP5 RCP45 ensemble mean (blue solid line) and RCP85 ensemble mean (red dashed line). The blue shaded area represents one standard deviation of 53 ensembles of RCP45. The linear decadal trends (denoted by ‘rc’) are listed in the box located at northwest side. All p values are less than 0.01. Note that most of RCP85 EM is overlapped by RCP45 EM before 2005. **b, c** Spatial pattern of decadal SST trends of CMIP5 RCP45 ensemble mean during 1970–1997 and 1999–2015 respectively, similar in RCP85. The black dots show the 99% significant level

Pacific during boreal winter (Fig. 9), which was not resolved by most of the CMIP5 models. Specifically, 43 out of 53 models were able to capture the warming during 1970–1997 (Fig. 11a); however, more than 40 models failed to simulate the observed warming slowdown in the SCS sea surface during 1999–2015 under the RCP4.5 scenario (Figs. 10, 11b).

The importance of the La Niña-like SST pattern in the eastern tropical Pacific during 1999–2010 was pointed out by Kosaka and Xie (2013), and then they successfully reproduced the so-called global “hiatus” phenomenon by artificially prescribing a cool SST tongue in the eastern tropical Pacific in a GCM. Compared to the remote atmospheric forcing, the effect of the oceanic circulation changes due to the Luzon Strait intrusion is rather small, accounting for 12% of the total SST variance. One of the reasons is that the SCS throughflow transport is about 4–6 Sv,

with local effect within the northern SCS shelf (Wei et al. 2016a). On the other hand, the SCS general circulation is anticyclonic as an oceanic response to the monsoon winds. The SCS throughflow is strengthened not only by the Luzon Strait intrusions, but also the enhanced EAWM. As shown in Fig. 4, the cooling rate on the northern SCS shelf due to the wind effect in the FSC run (Fig. 4d) is even larger than the oceanic boundary forcing in the FBC run (Fig. 4f). This finding indicates that for a regional model, increasing local ocean model resolution may help to better resolve the regional spatial patterns through improved bathymetry oceanic pathways, while improving large-scale atmospheric and oceanic boundary forcing can help to control the model long-term trend by linking local processes to large-scale remote forcings.

Fig. 11 SCS and eastern Tropical Pacific SST trends. **a** The SCS SST trends from 3 observations, RegCM-FVCOM model and 53 CMIP5 RCP45 runs during 1970–1997. **b** Similar with **a**, but for the trends during 1999–2015. **c**, similar with **b**, but for eastern Tropical Pacific. Each number of 1–53 corresponds to one simulation listed in Table 1. The solid filled bars represent the trends reach 90% significant level,

while the white blank bars are not. The grey dashed lines highlight those good CMIP5 members successfully capturing the slowdown of SST warming. **d–f** Frequency distribution of SST linear trend corresponding to **a–c** respectively. The stars indicate the trends distribution of observations (black stars), RegCM-FVCOM (blue star) and CMIP5 ensemble mean (orange star)

Table 1 List of CMIP5 simulations, corresponding to the number indices in Figs. 11, 13 and 14

01 ACCESS1-0_r1i1p1	19 CSIRO-Mk3-6-0_r3i1p1	37 inmcm4_r1i1p1
02 ACCESS1-3_r1i1p1	20 CSIRO-Mk3-6-0_r4i1p1	38 IPSL-CM5A-LR_r2i1p1
03 CCSM4_r1i1p1	21 CSIRO-Mk3-6-0_r5i1p1	39 IPSL-CM5A-LR_r3i1p1
04 CCSM4_r2i1p1	22 CSIRO-Mk3-6-0_r6i1p1	40 IPSL-CM5A-LR_r4i1p1
05 CCSM4_r3i1p1	23 CSIRO-Mk3-6-0_r7i1p1	41 IPSL-CM5A-MR_r1i1p1
06 CCSM4_r4i1p1	24 CSIRO-Mk3-6-0_r8i1p1	42 IPSL-CM5B-LR_r1i1p1
07 CCSM4_r5i1p1	25 CSIRO-Mk3-6-0_r9i1p1	43 MIROC-ESM-CHEM_r1i1p1
08 CCSM4_r6i1p1	26 GFDL-CM3_r1i1p1	44 MIROC-ESM_r1i1p1
09 CESM1-BGC_r1i1p1	27 GFDL-ESM2G_r1i1p1	45 MIROC5_r1i1p1
10 CESM1-CAM5_r1i1p1	28 GFDL-ESM2M_r1i1p1	46 MIROC5_r2i1p1
11 CESM1-CAM5_r2i1p1	29 GISS-E2-H-CC_r1i1p1	47 MIROC5_r3i1p1
12 CESM1-CAM5_r3i1p1	30 GISS-E2-H_r1i1p1	48 MPI-ESM-LR_r1i1p1
13 CMCC-CM_r1i1p1	31 GISS-E2-H_r1i1p2	49 MPI-ESM-LR_r2i1p1
14 CMCC-CMS_r1i1p1	32 GISS-E2-H_r1i1p3	50 MPI-ESM-LR_r3i1p1
15 CNRM-CM5_r1i1p1	33 GISS-E2-R-CC_r1i1p1	51 MPI-ESM-MR_r1i1p1
16 CSIRO-Mk3-6-0_r10i1p1	34 GISS-E2-R_r1i1p1	52 MRI-CGCM3_r1i1p1
17 CSIRO-Mk3-6-0_r1i1p1	35 GISS-E2-R_r1i1p2	53 NorESM1-M_r1i1p1
18 CSIRO-Mk3-6-0_r2i1p1	36 GISS-E2-R_r1i1p3	

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