Eddy-Induced Heat Transport in the South China Sea

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ABSTRACT: Eddy-induced heat transport (EHT) in the South China Sea (SCS) is important for the heat budget. However, knowledge of its variability is limited owing to discrepancies arising from the limitation of the downgradient method and uncertainties arising from numerical models. Herein, we investigated the spatiotemporal variability and dynamics of EHT using a well-validated assimilated model. In particular, to the southeast of Vietnam (SEV) and west of Luzon Strait (WLS), significant values of annual mean EHT are observed and most EHT is confined in the upper 400 m. EHT also exhibits significant seasonality, and the largest EHT amplitude in autumn at SEV is mainly driven by the wind stress curl, while that in winter at WLS is mainly related to the Kuroshio intrusion. Energy budget analysis reveals that both the barotropic and baroclinic instabilities increase the eddy kinetic energy in autumn at SEV, whereas only the barotropic instability contributes to the eddy kinetic energy at WLS in winter. Specially, an up-gradient EHT is observed at WLS in all four seasons, characterized by the same directions between EHT and mean temperature gradient. The upgradient EHT at WLS is induced by the baroclinic instability through an inverse energy transfer, which is generated by the interaction between the Kuroshio intrusion and topography below the surface layer. Moreover, the most significant upgradient EHT in winter shows a wave-like southwestward-propagating pattern in the subsurface layer.

KEYWORDS: Mesoscale processes; Ocean dynamics; Eddies; Data assimilation; Ocean models; Oceanic variability

1. Introduction

Heat transport induced by mesoscale eddies, namely, the eddy-induced heat transport (EHT), is crucial for the global heat budget. It is a salient contributor to the total heat transport, particularly for the meridional component of EHT (MEHT) (Roemmich et al. 2001). In some regions, MEHT can be greater than 20% of the total meridional heat transport (MHT) and significantly contributes to the variability of MHT (Sun et al. 2019). Thus, EHT has a significant impact on the climate and ecological systems (Macdonald and Wunsch 1996; Volkov et al. 2008).

Simultaneous current velocity and temperature observations are challenging to achieve; thus, EHT is a poorly observed quantity in the ocean. Therefore, two frequently used

methods have been used to study the spatiotemporal variability and dynamics of EHT: 1) downgradient method, which assumes that the heat flux has the opposite sign with the time-mean temperature gradient averaged over certain depth and 2) numerical models. Based on observed sea surface height (SSH) and climatological temperature data, Stammer (1998) estimated the global EHT with the downgradient method, which is the product of eddy diffusivity based on a mixing length argument and time-mean temperature gradient averaged over upper 1000 m (later on referred as EHT_{dg}). He found that EHT_{dg} is significant along the western boundary regions of oceans but insignificant in the interior basins. These findings are consistent with the general patterns derived from ocean current and temperature observations collected at different locations in the western Pacific and the Atlantic (Wunsch 1999). Later, using output from an eddy-resolving general circulation model, Jayne and Marotzke (2002) estimated EHT derived from the product of the velocity ($\mathbf{V}$) and the potential temperature (θ) anomalies (later on referred as $EHT_{\mathbf{V}\theta}$). Those anomalies are calculated by subtracting their time means during the entire 9-yr period of their study. Despite the similarity in the general pattern of globally estimated EHT using the two methods, significant differences are observed in the structure and magnitude of the derived EHT in some regions. On the one hand, the downgradient method failed to predict the $EHT_{\mathbf{V}\theta}$ structure, particularly in the equatorial region

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where the convergence of a large amount of $EHT_{V\theta}$ is found, as well as in the western boundary region where a smaller spread of northward $EHT_{V\theta}$ occurs. Instead, an upgradient location, which implies that EHT and time-mean temperature gradient have the same sign, was found in the equatorial area, where EHT is consistent with the observations. This upgradient feature of $EHT_{V\theta}$ was also described in another model study in the Kuroshio Extension, particularly below 200 m (Yim et al. 2010), suggesting that different mechanisms can be responsible for this upgradient feature in these regions. On the other hand, most models exclude constraints on observations; hence, EHT obtained from models can suffer from uncertainties arising from the subsurface (30–400 m) temperature and current velocity bias, particularly because the EHT is integrated in the vertical (Kuhlbrodt and Gregory 2012). Furthermore, a large bias in the simulated subsurface temperatures, which can be greater than 2°C (Chakraborty et al. 2015), can significantly affect the accuracy of calculated $EHT_{V\theta}$ because the vertical structure of $EHT_{V\theta}$ is associated with the subsurface temperature, as suggested by observations (Chen et al. 2012; Sun et al. 2019).

Such method-induced differences in EHT are also observed in regional studies conducted in the South China Sea (SCS), which is the largest semiclosed marginal sea in the northwest Pacific (Fig. 1). The SCS contains active mesoscale eddies that have a significant influence on the velocity and temperature in the upper 400 m (Wang et al. 2003; Chelton et al. 2007; Chen et al. 2011). As represented by the covariability between the velocity and temperature anomalies, the variability in EHT in SCS is comparable to that in the Kuroshio Extension (Qiu and Chen 2005; Chen et al. 2011; Pan and Sun 2018).

Previous studies have suggested noticeable discrepancies in the spatial pattern and magnitude between EHT_{dg} and $EHT_{V\theta}$ in SCS. Using the same downgradient method as used by Stammer (1998) and Chen et al. (2012) estimated EHT_{dg} in SCS using satellite altimetry data, Argo profiles, and climatological temperature data. Their results showed large northward EHT to the southeast of Vietnam (SEV) and southward EHT to the west of Luzon Strait (WLS). Additionally, the spatial distribution of EHT was mainly determined by the annual mean meridional temperature gradient. Moreover, a strong seasonality was observed. At WLS, a large amount of southward EHT was observed in winter, whereas at SEV, a large amount of northward EHT was found in summer. Such variability is related to baroclinic instability, which depends on the vertical velocity shear, latitude, and stratification.

Conversely, based on $EHT_{V\theta}$ obtained directly from an eddy-resolving ocean model, Wang (2011) and Jiang et al. (2016) found spatial distributions of EHT different from those found by Chen et al. (2012). In particular, a strong northward EHT was found along the western boundary currents, accompanied by a strong southward EHT on the right-hand side of the currents. Further, Wang (2011) reported the limits of using the downgradient method in the SCS by showing the same directions between annual mean EHT and annual-mean temperature gradient averaged over the whole water column.

FIG. 1. Study area. Black dots and lines denote the initial locations and subsequent trajectories of the individual Argo floats, respectively. The red star indicates the locations of three mooring buoys. The thick black line is the 200 m isobath.

This method-induced difference in SCS may indicate upgradient regions. It may be also associated with the model uncertainties arising from the subsurface ocean temperature and current velocity variability, which were not well validated in previous modeling studies; thus, the variability and dynamics of EHT remain unresolved. Therefore, in this study, EHT was investigated using an assimilated model that simulates reasonable temperature and current velocity structures in the vertical. Two questions were addressed: 1) what are the spatiotemporal variability and dynamics of $EHT_{V\theta}$ in SCS based on a well-validated assimilated model and 2) what are the spatial differences between $EHT_{V\theta}$ and EHT_{dg} in SCS and what are the possible reasons for these differences?

The remainder of this study is organized as follows. In section 2, the basic model configuration and assimilation scheme related to the study of EHT are presented. In section 3, the equations used for calculating $EHT_{V\theta}$ and EHT_{dg} are provided, along with the associated observational data. In section 4, the model validation concerning the vertical structure of the temperature and current velocity is shown. The EHT_{dg} -related fields from the model and observation are also compared. The modeled $EHT_{V\theta}$ variability and its dynamics are investigated in section 5. In section 6, the upgradient feature of EHT, which is the most significant difference between $EHT_{V\theta}$ and EHT_{dg} , is further discussed. Moreover, the contributions of $EHT_{V\theta}$ and EHT_{dg} to EHT are investigated by

separating EHT into rotational and divergent parts. The findings of this study are summarized in [section 7](#).

2. Model simulation

a. Model configuration

We applied the Massachusetts Institute of Technology general circulation model (MITgcm) ([Marshall et al. 1997](#)) to SCS for the area from 2° to 25°N and from 105° to 128°E . To adequately resolve the eddy field, the model had a horizontal resolution of $1/10^{\circ} \times 1/10^{\circ}$ ($\sim 10\text{ km} \times 10\text{ km}$) and 75 vertical levels in the z coordinate. Further, the model had a finer spacing in the subsurface layer; the vertical spacing above the depth of 200 m was less than 10 m to more adequately reflect the vertical structure of the mesoscale eddies. The topography used was derived and interpolated to the model grid from the global ETOP02v2 dataset provided by the National Oceanic and Atmospheric Administration ([NOAA/NGDC 2006](#)). We also used atmospheric forcing data comprising Cross-Calibrated Multi-Platform (CCMP) 6-hourly sea surface wind vector data [$1/4^{\circ} \times 1/4^{\circ}$ ($\sim 25\text{ km} \times 25\text{ km}$); [Atlas et al. 2011](#)] as well as the 6-hourly reanalysis sea surface heat flux, evaporation–precipitation, air temperature, and specific humidity data obtained from the U.S. National Centers for Environmental Prediction (NCEP; $1.875^{\circ} \times 1.875^{\circ}$; [Kalnay et al. 1996](#)). The different use of wind and buoyancy fluxes data can be justified by two reasons: 1) high-resolution wind data are important in simulating the perturbation processes such as the wave propagation and mesoscale eddies in the SCS, because these dynamical processes depend on the location and magnitude of the wind stress ([Wang et al. 2003, 2005](#)), and 2) the buoyancy fluxes in the SCS from different datasets share similar magnitude and spatial pattern (not shown). The initial temperature and salinity fields were obtained from the *World Ocean Atlas 2013* version 2 (WOA13; [Locarnini, et al. 2013](#)) climatological dataset, which has a horizontal resolution of $1/4^{\circ} \times 1/4^{\circ}$ and 33 vertical levels with finer spacing close to the sea surface. The daily temperature, salinity, and current velocity data at the open boundaries of SCS (the area north of Taiwan and the western Pacific) were derived from the Hybrid Coordinate Ocean Model (HYCOM; [Bleck 2002](#)) reanalysis product.

The horizontal mixing scheme was obtained from the biharmonic friction scheme ([Smagorinsky 1963](#)), and the corresponding coefficient was in the range of $10^5\text{--}10^8\text{ cm}^2\text{ s}^{-1}$ which is calculated based on the grid length and a deformation rate associated with horizontal tension and shear of velocities ([Adcroft et al. 2018](#), section 2.21.1.3). The vertical turbulence parameterization scheme was the nonlocal K -profile parameterization (KPP) that separates the vertical water column into the upper boundary and interior layers based on the differences in dynamical factors. KPP used herein was a modification of the parameterization used by [Large et al. \(1994\)](#) and [Durski et al. \(2004\)](#) and was previously successfully applied to SCS ([Luo et al. 2016](#)).

The model was spun up for 10 years from zero velocities and climatological values of temperature and salinity, as well as

climatological atmospheric forcing till the integrated eddy kinetic energy (EKE) in SCS was stable. Thereafter, the simulation was started based on the initial conditions produced by the spinup. The model was integrated from 2012 to 2015 to derive the daily temperature, potential density, SSH, and current velocity data that could be used for studying EHT in SCS.

b. Assimilation scheme

To reduce the subsurface temperature error and improve the accuracy of calculated EHT, a joint assimilation scheme for the sea surface temperature (SST) and SSH was used in the assimilation module of MITgcm. SST and SSH data used for the assimilation were obtained from the Advanced Very High-Resolution Radiometer (AVHRR; [Casey et al. 2010](#)) and European Copernicus Marine Environment Monitoring Service (CMEMS). The horizontal and temporal resolutions for both SST and SSH were $1/4^{\circ} \times 1/4^{\circ}$ and daily, respectively.

The nonsolar heat flux was used to correct the SST to maintain the dynamic relation between the surface heat flux and simulated temperature in the boundary layer because the solar heat flux at the ocean surface should not depend on the modeled SST ([Barnier et al. 1995](#); [Tang et al. 2001](#)). The nonsolar heat flux refers to the heat flux without solar short-wave radiation, which includes the longwave radiation and latent and sensitive heat flux. In MITgcm, the vertical turbulence approximation of KPP defined a boundary layer that is dominated by nonsolar heat flux and momentum flux and an interior layer that is affected by vertical shear and internal waves. In the boundary layer, we established a relation between the nonsolar heat flux and SST error using a 5-day-weighting nudging method. Here, we assumed that the nonsolar heat flux bias (ΔQ) has a major effect on the simulated SST error between the simulated SST and observed SST (ΔT), similar to [Barnier et al.'s \(1995\)](#) study. To establish a relation between ΔQ and ΔT , we assume that ΔQ originates from ΔT . After establishing the relation between the required correction of SST and nonsolar heat flux, we adjusted the temperature in the upper boundary layer to improve the accuracy of the upper boundary layer depth associated with the turbulence parameterization scheme.

Moreover, the vertical structure of the temperature below the boundary layer was improved when we established an SSH correction based on the conservation of the potential vorticity (PV) ([Cooper and Haines 1996](#)); this was performed to ensure that the assimilated output was dynamically consistent with the model in the ocean interior. The dynamical SSH projection ([Cooper and Haines 1996](#)) was realized by establishing a vertical displacement (ΔZ) using the PV conservation hypothesis. ΔZ was calculated from the bias of SSH anomalies between the simulated and observed SSH anomalies. Because the adiabatic condition and PV conservation hypothesis were not satisfied in the boundary layer, we applied ΔZ only below the boundary layer depth. This SSH projection method has been successfully employed in the state-of-the-art ocean reanalysis system [ORAS4 ([Balmaseda et al. 2013](#)); HYCOM ([Chassignet et al. 2007](#))]. In particular, HYCOM reanalysis showed a reasonable performance in

simulating the intraseasonal variability in the Kuroshio intrusion in SCS (Zhang et al. 2017; Yang et al. 2019).

The nonphysical temperature source/sink originating from the temperature adjustment was evaluated because it may affect the estimated EHT. Because the surface heat flux adjustment would not affect the dynamical balance of the temperature equation in the ocean interior, we examined the interior temperature adjustment from the SSH assimilation. This temperature source term, which is represented by the vertical replacement of the bias between the simulated and observed SSH (Cooper and Haines 1996), involves the evolution of temperature tendency. The monthly results of these two terms suggested that the temperature source term was considerably smaller than the temperature tendency (Fig. S1 in the online supplemental material), particularly above the boundary layer depth. Considering that the largest EHT from model was mainly located above the boundary layer depth (not shown), we concluded that the correction term from the assimilation was relatively small in the temporal evolution of temperature tendency and did not significantly affect the simulated EHT in this study. Using this assimilation scheme, we reduced the root-mean-square error (RMSE) in the subsurface temperature from 2.9°C in the dynamic model to 1.3°C in the reanalysis system. The validation of EHT in SCS is presented in section 4.

3. Calculation of EHT

a. Direct evaluation of $EHT_{V\theta}$ from the model

In this study, we focused on the time scale of mesoscale eddies. Consequently, the time mean and deviation of the velocity and temperature were different from those reported in the study by Jayne and Marotzke (2002) considering that their EHT calculation included all perturbations because they used the time mean for the entire period of their study. To derive EHT corresponding to the mesoscale eddies, the anomalies in horizontal velocity ($\mathbf{V}'$) and potential temperature (θ') were first derived by subtracting the climatological seasonal means ($\bar{\mathbf{V}}$ and $\bar{\theta}$) from $\mathbf{V}$ and θ . Then, the time series of these anomalies ($\mathbf{V}'$ and θ') were high-pass filtered with 100 days because most eddy energy is contained within a period of less than 100 days (Volkov et al. 2008). Thus, the estimated EHT included most eddy variability, including the variability attributed to the Ekman process. The time-mean $EHT_{V\theta}$ directly evaluated from the mesoscale-related velocity and temperature anomalies was calculated for each model grid in 2012–15 as follows:

$$EHT_{V\theta}(x, y) = \int_{-h}^0 \rho c_p \overline{\mathbf{V}'\theta'} dz = \int_{-h}^0 \rho c_p \overline{(\mathbf{V} - \bar{\mathbf{V}})(\theta - \bar{\theta})} dz, \quad (1)$$

where the overbars denote the time mean over a certain period, which is the annual or seasonal mean. The prime on the variables indicates a deviation from the climatological seasonal mean. The zonal EHT (ZEHT) and MEHT were calculated using the two components of the horizontal velocity ($\mathbf{V}$), namely, u and v . Other terms in Eq. (1) include seawater density ρ , specific heat capacity of seawater at a constant

pressure c_p , which has a value of $4189 \text{ J kg}^{-1} \text{ } ^\circ\text{C}^{-1}$, potential temperature θ , and the integrated depth in the vertical direction h .

b. Calculation of EHT_{dg} using the downgradient method

Herein, the $EHT_{V\theta}$ values calculated using the model were compared with the EHT_{dg} values calculated using the downgradient method [Eq. (2)]. Additionally, the zonal (meridional) EHT_{dg} was determined using $-\int_{-h}^0 \rho c_p \kappa \partial\theta/\partial x dz$ ($-\int_{-h}^0 \rho c_p \kappa \partial\theta/\partial y dz$). As suggested by this equation, the zonal (meridional) part of EHT is always in the downgradient direction, i.e., its sign is always opposite to the direction of the zonal (meridional) part of the time-mean depth-integrated temperature gradient because the eddy diffusivity κ is positive.

$$EHT_{dg}(x, y) = - \int_{-h}^0 \rho c_p \kappa \nabla_h \bar{\theta} dz, \quad (2)$$

where $\bar{\theta}$ is the time-mean temperature at every model depth. The eddy diffusivity κ was derived using the following equation:

$$\kappa = 2\alpha K_E T_{alt},$$

$$T_{alt} = \frac{1}{C_\xi(0)} \int_0^{T_0} C_\xi(\tau) d\tau, \quad (3)$$

where the value of the mixing efficiency (α) is 0.01 (Chen et al. 2012). K_E is near-surface EKE. In the eddy-mixing time scale (T_{alt}) equation, C_ξ is the autocovariance of the SSH time series, and T_0 is the first zero-crossing point of C_ξ .

To justify the comparison between EHT_{dg} and $EHT_{V\theta}$, the EHT_{dg} -related fields, namely, $\int_{-h}^0 \nabla_h \bar{\theta} dz$, K_E , and T_{alt} , were calculated. The horizontal temperature gradient was calculated using the monthly temperature climatology from the model and in situ observation, which is obtained from WOA13. Furthermore, K_E and T_{alt} were computed from the daily SSH from the model and satellite altimetry data. Note that K_E was calculated from the surface velocity anomalies based on SSH and geostrophic approximation. To extract the mesoscale variability, the same process of deriving the velocity anomalies described in section 3a is applied to SSH, which is also used to calculate K_E and T_{alt} .

4. Validation of modeled eddy-related variability in SCS

The variability in EHT is associated with the vertical temperature and velocity anomalies, particularly above 400 m (Chen et al. 2012); thus, the corresponding temperature profiles from the Argo floats and current velocity profiles from the in situ mooring buoys were used to validate the model results and ensure reliable analysis. Moreover, the surface EKE and depth-integrated temperature gradient were validated to justify the comparison between $EHT_{V\theta}$ and EHT_{dg} .

a. Temperature profiles from Argo floats in SCS

Because the dataset provided by the Argo floats was large and included the vertical temperature profiles of the mesoscale

FIG. 2. Monthly RMSEs in the vertical profile of the temperature and temperature anomaly: (a),(b) monthly RMSE and (c),(d) the corresponding averaged RMSEs over 12 months.

eddies, the daily temperature data obtained from the Global Argo Data Repository were compared with the daily temperature data obtained using the model. The Argo floats cover most parts of SCS, where there is a high probability of eddy activities (Fig. 1), and provides 5229 temperature profiles for the simulation period (2012–15) with at least 310 samples in each month. This ensured that the validation was statistically reliable. To obtain RMSE, the daily temperature or temperature anomaly of the Argo profiles during the simulation period of 2012–15 was grouped into each month of a calendar year. In each month, the monthly RMSE was calculated at every standard depth between the Argo and model data, which corresponded to the Argo data at the same times and locations. Note that the daily temperature anomaly of the Argo profiles was obtained by removing the climatological monthly mean temperature, which was achieved by interpolating the monthly WOA13 temperature to the corresponding Argo profile locations and times. This approach is similar to the process used by Qiu and Chen (2005).

Figure 2 shows the depth–month RMSE of the temperature and temperature anomaly profiles between the Argo and model data. To demonstrate the improvement in the assimilated temperature, the RMSE of the unassimilated results were also obtained (SI, Fig. S2). Generally, the model showed a significantly larger bias before assimilation than that after assimilation (Fig. 2). The averaged RMSE, which refers to the average of monthly RMSEs, shows that the largest RMSE ($\sim 3^{\circ}\text{C}$) occurs at a depth of 100 m (Fig. S2c). Further, the RMSE results of the monthly temperature profiles show that

the error significantly increased from July to December, particularly at depths greater than 50 m (Fig. S2a). The RMSE of the temperature anomaly, which is closely related to the EHT calculation, is also shown in the figure. The overall averaged RMSE of the temperature anomaly before assimilation was approximately 1.5°C , and the maximum error ($\sim 2^{\circ}\text{C}$) was observed at a depth of approximately 80 m (Fig. S2d). Moreover, the monthly RMSE of the temperature anomaly was large, i.e., between 50 and 150 m, and the maximum error occurred during late winter (January and February) (Fig. S2b). Overall, the averaged RMSE of the temperature and temperature anomaly was large, particularly from 50 to 150 m. The error increased later in the year, indicating that the large temperature bias in the unassimilated model significantly influenced the mean value and variability results of the calculated EHT.

A marked improvement was observed in the modeled temperatures after assimilation (Fig. 2). The overall annual mean RMSE of both the temperature and temperature anomaly was less than 1°C , which was half the RMSE value reported by Chakraborty et al. (2015). The largest RMSE was reduced to $\sim 1.3^{\circ}\text{C}$ (Figs. 2c,d), which was less than half the magnitude of the error before assimilation. Additionally, the monthly RMSE (Figs. 2a,b) values showed that the maximum temperature bias at depths of 50–150 m was less than 1.5°C . More importantly, the seasonality in the error almost disappeared, indicating that the assimilated temperature and temperature anomaly were dynamically adjusted to reasonable values. Considering that there was an error of 0.6°C in the satellite SST dataset itself, the results of the assimilated model likely

FIG. 3. (a)–(f) Time–depth profiles of the zonal component of velocity anomalies of three mooring buoy stations based on the (left) observation and (center) model. (g)–(l) As in (a)–(f), but for the meridional component of velocity anomalies. Unit of velocity is m s^{-1} . (right) The corresponding RMSE and correlation between the observation and model at each depth.

yielded a more robust vertical temperature structure for SCS than the unassimilated model.

b. Current velocity profiles from the three mooring buoys in the northwest of SCS and at WLS

In addition to the validation of the temperature variability, three mooring buoys with daily current velocity data were collected to validate the eddy-related velocity variability based on the model. Two mooring buoys were located at WLS (stations 1 and 2, Fig. 1), where significant EKE and EHT were located (Zhuang et al. 2010; Chen et al. 2012). The other mooring buoy was located northwest of SCS (station 3, Fig. 1), where a strong surface EHT was located, as indicated by the observation (Pan and Sun 2018). The velocity anomaly was achieved by subtracting the time mean from the velocity during each observational period.

Figure 3 presents the time–depth profiles of the zonal and meridional components of the velocity anomalies from the observation and model. As expected, the velocity anomaly at WLS (stations 1 and 2) exhibited a magnitude (e.g., October–November in Fig. 3a) larger than that at the northwest of SCS (station 3). Moreover, a comparison between the observation and model results showed that the model well simulated the magnitude and variability of the velocity anomaly. This velocity anomaly validation between the observation and model can be seen more clearly from the RMSE and the correlation between them at each depth. A relatively larger bias of the velocity anomaly ($\sim 0.08 \text{ m s}^{-1}$) was located at the top 100 m for stations 1 and 2, where the largest magnitude of the velocity anomaly was located. A high correlation (>0.8) of the time series of the velocity anomaly between the observation and model was observed at the same depth layer, indicating that

FIG. 4. Annual mean surface EKE (color shading; $10^6 \text{ cm}^2 \text{ s}^{-2}$) and depth-integrated temperature gradient in the top 400 m (vectors; $10^4 \text{ }^\circ\text{C m}^{-1}$) from the (a) observation and (b) model.

this model reasonably simulated the mesoscale variability of the current velocity, particularly at places with significant eddy activities.

c. Climatological surface EKE and temperature gradient in SCS

The structure and magnitude of the variables determining the EHT_{dg} was validated to justify the difference between $\text{EHT}_{\text{v}\theta}$ and EHT_{dg} . According to Eq. (2), EHT_{dg} was mainly determined by the surface EKE and depth-integrated temperature gradient. Therefore, these two terms obtained from the model were validated against those obtained from the observation (Fig. 4). The depth-integrated temperature gradient over the top 400 m was derived in this section for comparison with the result achieved by Chen et al. (2012).

Results indicated that the model exhibited similar spatial patterns of the surface EKE and depth-integrated temperature gradient as those obtained from the observations, although EKE of the model was slightly smaller than that of the observation. Two important regions with a significant surface EKE were observed at WLS and SEV using both the observation and model. Moreover, these two regions showed a large temperature gradient but with opposite signs in the meridional direction, which was positive and negative at WLS and SEV, respectively. Based on Eq. (2), this finding indicates that the meridional EHT_{dg} was negative and positive at WLS and SEV, respectively. This result was consistent with that obtained by Chen et al. (2012). An obvious difference in the temperature gradient between the model and observation was observed west of Luzon island at approximately 12°N . This occurrence is likely related to different averaging periods, which were 1955–2012 for *WOA13* and 2012–15 for the

model in this study. Overall, our model can reproduce the EHT_{dg} pattern, ensuring a reliable comparison between EHT_{dg} and $\text{EHT}_{\text{v}\theta}$.

5. Modeled $\text{EHT}_{\text{v}\theta}$ in SCS

The annual and seasonal means of $\text{EHT}_{\text{v}\theta}$ of the entire SCS derived from the model [Eq. (1)] were analyzed. Before this step, it was necessary to obtain a general impression of the eddy activity intensity and the covariability between the temperature and velocity anomalies in SCS. The depth integral of EKE obtained from the model was calculated as

$$\text{EKE}(x, y) = \frac{1}{2} \int_{-h}^0 (u'^2 + v'^2) dz, \quad (4)$$

where the velocity anomalies only included signals corresponding to a time scale of less than 100 days, which were consistent with the EHT calculations.

The contribution of the covariability between $\mathbf{V}'$ and θ' to the time-mean $\text{EHT}_{\text{v}\theta}$ can be decomposed further in the frequency domain and represented by the integral of the co-spectrum of $\mathbf{V}'$ and θ' , which is equal to the real part of their cross spectrum (Bryden 1979):

$$\begin{aligned} \text{EHT}_{\text{v}\theta}(x, y) &= \frac{1}{\tau} \int_{-h}^0 \int_{\tau}^0 \rho c_p \mathbf{V}' \theta' dt dz \\ &= \frac{1}{\tau} \int_{-h}^0 \int_{\tau}^0 \rho c_p \text{real}(\hat{\mathbf{V}} \hat{\theta}^*) df dz, \end{aligned} \quad (5)$$

where τ is the averaging period, $\hat{\mathbf{V}}$ is the Fourier transform of the velocity anomaly $\mathbf{V}'$, $\hat{\theta}^*$ is the complex conjugate of the

FIG. 5. (a) Depth-integrated EKE in SCS (shading) and the cospectra of ZEHT and MEHT (red and blue contour lines, respectively; values larger than 80 MW m^{-1} are plotted). (b) Area-averaged cospectra of ZEHT (red line) and MEHT (blue line) in the dashed boxes at SEV and WLS. The dashed boxes in (a) denote the regions with the largest values of the EHT cospectra at SEV and WLS.

Fourier transform of the temperature anomaly θ' , and f denotes the frequency in this equation.

The most significant feature of the depth-integrated EKE was found at SEV (Fig. 5a), and there was a secondary feature located at WLS (Fig. 5a). These findings are similar to the results derived from the satellite altimetry product (Wang et al. 2005). As expected, the locations with the most significant cospectra of ZEHT and MEHT (red and blue lines in Fig. 5a, respectively) coincide with the EKE locations, with the magnitudes of the cospectra at SEV being larger than those at WLS (Fig. 5b). Moreover, the most significant values of the cospectra of ZEHT and MEHT at SEV and WLS were in the frequency band of 32–100 days, whereas the largest values of the cospectra of EHT were in the frequency of 70 and 55 days at SEV and WLS, respectively (Fig. 5b). The covariance period of EHT at WLS was close to the annual mean period estimate (~ 60 days) of the eddy shedding in the same area based on the mean velocity of the Kuroshio velocity, width of the Luzon Strait, and empirical constant (Wu and Chiang 2007).

a. Annual mean $\text{EHT}_{\mathbf{v}_\theta}$

To investigate the modeled $\text{EHT}_{\mathbf{v}_\theta}$ of the entire SCS, the annual zonal and meridional means of $\text{EHT}_{\mathbf{v}_\theta}$ at each point in the model grid were integrated over the entire water column [according to Eq. (1)] and presented in Fig. 6. Generally, significant values of both ZEHT and MEHT were found at SEV and WLS, where the largest EKE values and EHT cospectra are located. Furthermore, a strong, dipole-like pattern

in ZEHT and MEHT was found at SEV (Figs. 6a,b), indicating that the rotational part of EHT dominated in this region. Additionally, ZEHT results indicated that the westward EHT was larger at WLS ($\sim 35 \text{ MW m}^{-1}$) than at SEV ($\sim 20 \text{ MW m}^{-1}$), whereas MEHT results showed that the northward EHT was larger at SEV ($\sim 45 \text{ MW m}^{-1}$) than at WLS ($\sim 30 \text{ MW m}^{-1}$).

Overall, the spatial patterns in ZEHT and MEHT were consistent with the modeling results shown by Wang (2011). However, the southward MEHT along the western boundary area was less prominent in our study, likely because our estimates only considered the short-term perturbation, i.e., the mesoscale eddy time scale (Fig. S3). Compared with the EHT_{dg} results obtained using the downgradient method, the modeled $\text{EHT}_{\mathbf{v}_\theta}$ results produced different spatial patterns in ZEHT and MEHT at SEV and WLS, respectively. Our model can reproduce the annual mean state of EHT_{dg} similar to that obtained from the observation (section 4c); therefore, the difference between $\text{EHT}_{\mathbf{v}_\theta}$ and EHT_{dg} indicates that the downgradient method may not predict $\text{EHT}_{\mathbf{v}_\theta}$ correctly. For qualitative examination of the upgradient and downgradient locations of EHT in SCS, the relation between the zonal (meridional) temperature gradient and zonal (meridional) $\text{EHT}_{\mathbf{v}_\theta}$ obtained from the model was evaluated based on the downgradient assumption (section 3b). The annual mean meridional temperature gradient (contour lines in Fig. 6b) showed that the sign of MEHT is the same as that of the meridional temperature gradient at WLS, implying that EHT was in the upgradient direction in this region.

FIG. 6. Annual means of the depth-integrated (a) ZEHT and (b) MEHT in SCS. (c) Meridionally integrated ZEHT and (e) its integrated values over four different depth layers: the top 30 m, 30–400 m, 400–1000 m, and >1000 m. (d),(f) As in (c) and (e), but for zonally integrated MEHT. The contour lines in (a) and (b) denote the annual means of the depth-integrated zonal and meridional temperature gradient, respectively. Dashed (solid) lines denote negative (positive) values. The units of the temperature gradient are $10^3 \text{ } ^\circ\text{C m}^{-1}$.

The spatial distribution of $\text{EHT}_{\text{V}\theta}$ was further investigated with respect to the zonal integral of MEHT and the meridional integral of ZEHT. As shown in Fig. 6c, the meridionally accumulated ZEHT was directed westward at all longitudes, with the largest EHT [~ -9.5 TW (1 TW = 10^{15} W)] located at SEV

at approximately 110.5°E and the second-largest EHT (~ -7.8 TW) located at WLS at approximately 109°E . Conversely, the values of the zonally accumulated MEHT showed northward EHT at almost all latitudes (Fig. 6d). Furthermore, two locations with notable EHT of ~ 11.8 and ~ 13.5 TW were observed

FIG. 7. Seasonal means of the depth-integrated ZEHT in (a) spring, (b) summer, (c) autumn, and (d) winter. (e)–(h) As in (a)–(d), but for MEHT. (i)–(l) The magnitude (shading) and direction (arrows) of EHT. Contour lines denote the seasonal temperature gradients of ZEHT and MEHT. Dashed (solid) lines denote negative (positive) values. The units of the EHT and temperature gradient are MW m^{-1} and $10^3 \text{ }^\circ\text{C m}^{-1}$, respectively.

at 14°N , and 22°N , respectively. Possibly, the area with a strong northward EHT at approximately 10°N was partly canceled out by the southward EHT at the same latitude. Generally, large amounts of EHT are observed in regions where the currents are significantly variable; this variability can be associated with meanders in jet streams, otherwise known as baroclinic current instability (Rossby 1987; Chen et al. 2012).

The meridionally integrated ZEHT and zonally integrated MEHT were further divided into four layers: surface (0–30 m), subsurface (30–400 m), middepth (400–1000 m), and deep ocean (>1000 m) layers. Generally, EHT was confined to the upper 1000 m, which was consistent with the results achieved from the Argo floats and current-meter data (Chen et al. 2012; Wunsch 1999). Most of the horizontal EHT was found in the subsurface layer (Figs. 6e,f), suggesting that EHT dynamics

were confined to the upper-ocean part, which agreed with the results of previous studies (Böning and Cox 1988; Jayne and Marotzke 2002; Qiu and Chen 2005). Moreover, EHT in the subsurface layer mainly determined the spatial variation in the ZEHT and MEHT. The EHT in the surface layer, which was associated with the Ekman variability, only modulated the local ZEHT and MEHT. An exception to this was ZEHT in the middepth layer (Fig. 6e; dotted–dashed line), which made a small but noticeable contribution to EHT at almost all longitudes, indicating the important role of turbulent heat transport below the thermocline.

b. Seasonal means of $\text{EHT}_{\mathbf{V}_\theta}$

The annual mean $\text{EHT}_{\mathbf{V}_\theta}$ values showed large amounts of EHT in regions with strong currents that also exhibited

FIG. 8. Seasonal distributions of ZEHT and MEHT (solid lines) in the top 400 m along with the corresponding temperature gradients (dashed lines) at SEV and WLS. Green, red, black, and blue denote spring, summer, autumn, and winter, respectively. The unit of EHT is MW m^{-1} .

significant seasonal variations (Su 2005; Yuan et al. 2006). The seasonal variability in ZEHT and MEHT can be determined from the seasonal means of spring (March–May), summer (June–August), autumn (September–November), and winter (December–February). Generally, the locations of large seasonal EHT values coincided with the locations of high annual mean values; however, the magnitude of EHT significantly varied with the season (Figs. 7i–l). In particular, EHT at SEV reached its maximum and minimum in autumn ($\sim 60 \text{ MW m}^{-1}$) and spring ($\sim 30 \text{ MW m}^{-1}$), respectively, while EHT at WLS reached its maximum and minimum in winter ($\sim 55 \text{ MW m}^{-1}$) and summer ($\sim 25 \text{ MW m}^{-1}$), respectively.

Both ZEHT and MEHT exhibited a dipole-like structure at SEV; here, the maximum values were observed in autumn (Figs. 7c,g) and manifested as a feature with an anticyclonic circulation (Fig. 7k). In winter, ZEHT and MEHT exhibited striped maxima at WLS (Figs. 7d,h), and the circulation pattern was similar to that at SEV but covered a larger area (Fig. 7l). We also compared the seasonal values of ZEHT and MEHT with their time-mean depth-integrated temperature gradients. Similar to the mean value results, the upgradient MEHT mainly appeared at WLS in spring, autumn, and winter and the upgradient ZEHT appeared at the same location in winter. The region near SEV at approximately 112°E also corresponded to an upgradient EHT area in the regions with eastward and southward EHT in almost all seasons.

Figure 8 shows the vertical distributions of ZEHT and MEHT at SEV and WLS (within the areas marked by the dashed boxes in Fig. 5a). To separate the eastward and westward components of ZEHT, the eastward (westward) EHT was obtained by masking the values of westward (eastward) EHT in the same area and time. The same procedure was also performed for the northward and southward components of MEHT. Overall, significant seasonality was observed in ZEHT and MEHT at depths of 50–200 m, with the largest EHT located at a depth ($\sim 55 \text{ m}$) at SEV shallower than that at WLS ($\sim 100 \text{ m}$). The exception was the northward EHT at SEV, where the largest EHT occurred at a depth of $\sim 90 \text{ m}$. At SEV, the eastward and westward ZEHT (Figs. 8a,e, respectively) showed comparable magnitudes ($\sim 0.15 \text{ MW m}^{-1}$), while the amplitude of the northward MEHT (0.32 MW m^{-1}) (Fig. 8b) was approximately twice that of the southward MEHT (Fig. 8f). At WLS, a large contrast was observed between the westward and eastward ZEHT as well as between the northward and southward MEHT. In particular, both the westward and northward EHT (Figs. 8d,g), which exhibited magnitudes of $\sim 0.28 \text{ MW m}^{-1}$, were at least twice the magnitude of their counterparts (Figs. 8c,h).

Consistent with the depth-integrated EHT at SEV, the northward and westward EHT (Figs. 8b,e) were almost in the downgradient direction at all depths, whereas the eastward and southward EHT (Figs. 8a,f) were in the upgradient direction at

FIG. 9. Seasonal means of (a)–(d) squared buoyancy frequency (N^2) and (e)–(h) squared velocity shear averaged over 30–400 m. (i)–(l) Wind stress (vectors) and wind stress curl (shading).

all depths. At WLS, the westward and southward EHT (Figs. 8g,h) were directed downgradient, whereas the eastward and northward EHT (Figs. 8c,d) were directed upgradient, particularly below 50 m. One exception was that even though the depth-integrated ZEHT in winter exhibited a significant upgradient feature at approximately 116°E near WLS (Fig. 7d), its vertical structure, particularly the westward EHT (Fig. 8g), did not show such an upgradient feature with its zonal temperature gradient. This is because the negative mean zonal temperature gradient was canceled by the strong positive value at WLS at approximately 120°E as a regional mean.

c. Mechanisms that determine the variability of $EHT_{\nabla\theta}$

A significant seasonality in EHT at SEV and WLS was revealed in the last section. In this section, the possible driving forces for such a unique seasonal feature of EHT are investigated. Moreover, notable upgradient and downgradient EHT

were identified at these two regions. Its dynamic process is examined using an eddy energy budget analysis.

1) POSSIBLE DRIVING FORCES FOR THE SEASONAL VARIATION IN EHT

Previous studies suggested several driven forces which are responsible for the seasonal variability of the mesoscale eddies in the SCS. At the SEV, the local wind stress curl is considered to be the main reason for the seasonality of mesoscale eddy (Wang et al. 2006; Gan and Qu 2008). At the WLS, the Kuroshio intrusion and its interaction with wind and topography also provides energy and PV flux which favor the generation and variability of mesoscale eddy (Zhuang et al. 2010; Sheu et al. 2010). Therefore, these two factors are considered in this study to investigate the possible driving forces for the seasonal variability in EHT. In particular, the wind stress curl was calculated from the wind data obtained from the model input atmospheric forcing. The Kuroshio

intrusion was investigated using the zonal velocity shear [Eq. (6)] and buoyancy frequency [Eq. (7)] obtained from the Richardson number, a measure of baroclinic instability (Stammer 1998). For example, smaller Richardson number means larger baroclinic instability, which is accompanied by large mean flow shear and small ocean stratification. Note that the velocity shear and buoyancy frequency were averaged over 30–400 m because the largest EHT was located in this layer:

$$\left| \frac{\partial u}{\partial z} \right| = \frac{g}{\rho_0 f} \frac{\partial \rho}{\partial y}, \quad (6)$$

$$N = \sqrt{-\rho_0^{-1} g \partial \rho / \partial z}. \quad (7)$$

Coinciding with the locations of the largest EHT, all three factors showed a significant seasonality at SEV and WLS (Fig. 9). The large EHT at SEV in summer (Fig. 7j) can be associated with the large velocity shear and wind stress curl at the same area and time (Figs. 9f,j). As shown by previous studies (Chu et al. 1998; Wang et al. 2006), the significant lagged correlation (about 40–50 days) between local wind stress curl and SSH supports a westward Rossby wave mechanism, which is considered to be a possible candidate for the generation and maintaining of the eddy at SEV from summer to autumn. Meanwhile, a significant velocity shear in the subsurface layer occurred in autumn (Fig. 9g), which is consistent with former study that this eddy caused large variations in the thermocline structure (Chen et al. 2010). Then, this velocity shear weakened (Fig. 9h) as the cold northeasterly prevailed (Fig. 9l), causing a stronger mixing (Fig. 9d). In other words, the wind stress curl was the major factor for the seasonality in EHT at SEV. Because the velocity shear was analyzed in the subsurface layer, the role of Ekman in the EHT seasonality was also evaluated (Fig. S4). The result showed that the Ekman component of EHT exhibited larger seasonal variability than its mean state. Because of the seasonal reversal of wind stress (Fig. 9), the mean Ekman transport is partly canceled. But magnitude of Ekman component was considerably weaker than that of EHT (Figs. 6c,d), indicating that the velocity and temperature fluctuations in the Ekman layer (~25 m) were not significant compared with those in the subsurface layer (Figs. 6e,f). This finding was also consistent with EHT profiles as estimated from observation (Chen et al. 2012).

The dynamic background at WLS is more complicated, particularly in winter when the maximum EHT occurs (Fig. 7l). The Kuroshio intrusion, which was associated with the large velocity shear (Fig. 9h) and small buoyancy frequency (Fig. 9d), was likely the major factor inducing the large EHT in this area in winter. As for other seasons, the Kuroshio intrusion is relatively weak and wind stress curl also contributes to the eddy variability. The Kuroshio intrusion at WLS was extensively studied, and its seasonal variation was related to the eddy activity in this area (Wu and Chiang 2007; Liang et al. 2008; Hsin et al. 2008).

2) ANALYSIS OF THE EDDY ENERGY BUDGET

The role of perturbation fields in the mean flow is important to address the question of why EHT is directed downgradient

at SEV and upgradient at WLS. The four-box energy diagram of Lorenz (1955) showed the regional energy budgets, elucidating the role of fluctuating fields in the mean flow. Although a complete energy budget is beyond the scope of the present study, some insights can be gained by considering the energy transfer between mean and eddy energy, particularly for the baroclinic and barotropic instabilities. As performed in previous studies, a regional energy budget analysis for SCS that examines the seasonal production of turbulent kinetic energy was conducted (Böning and Budich 1992; Beckmann et al. 1994; Zhuang et al. 2010).

The eddy energy terms were EKE [Eq. (4)] and eddy potential energy [EPE; Eq. (8)], and the eddy energy-conversion terms were the baroclinic conversion term [T2; Eq. (9)] and barotropic conversion term [T4; Eq. (11)]. A positive value of T2 indicates that the mean potential energy is converted to EPE through baroclinic instability, and a part of EPE is converted to EKE through T3 [Eq. (10)]. A positive value of T4 indicates that the kinetic energy of the mean flow is converted to EKE through barotropic instability. The T3 term involving the vertical velocity is strongly modified by the upwelling and downwelling caused by the flow over rough bottom topography, which could mask a coherent signal. Therefore, only the T2 and T4 conversion terms were used to diagnose the baroclinic and barotropic instabilities. Note that all time-varying variables contained high-frequency variations with periods of less than 100 days, which were same as those for the other time-varying variables discussed in this study:

$$\text{EPE}(x, y) = \frac{g}{2\rho_0} \int_{-h}^0 \frac{\bar{\rho}^2}{d\bar{\rho}/dz} dZ, \quad (8)$$

$$\text{T2}(x, y) = \frac{g}{\rho_0} \int_{-h}^0 \frac{\overline{u' \rho'} (\partial \bar{\rho} / \partial x) + \overline{v' \rho'} (\partial \bar{\rho} / \partial y)}{d\bar{\rho}/dz} dZ, \quad (9)$$

$$\text{T3} = -\frac{g}{\rho_0} \int_{-h}^0 \overline{\rho' w'} dZ, \quad (10)$$

$$\text{T4}(x, y) = -\int_{-h}^0 \left[\overline{u' u'} \frac{\partial \bar{u}}{\partial x} + \overline{u' v'} \left(\frac{\partial \bar{v}}{\partial x} + \frac{\partial \bar{u}}{\partial y} \right) + \overline{v' v'} \frac{\partial \bar{v}}{\partial y} \right] dZ, \quad (11)$$

where ρ_0 is the reference seawater density and $\bar{\rho}$ is the horizontally averaged time-mean density.

Figure 10 shows the seasonal distributions of the depth-integrated EPE, EKE, T2, and T4. Similar to the maps of the EHT seasonal means (Figs. 7a,b), the largest values of EPE and EKE were located at SEV and WLS (Figs. 10a–h). However, their amplitudes significantly varied with the season. Generally, EPE was considerably larger at WLS than at SEV; further, the largest values were observed at WLS and SEV in winter and autumn, respectively. As indicated by a previous analysis, the main energy input at these two areas can be through the time-mean wind and buoyancy forcing. Conversely, EKE was larger at SEV than at WLS, with the largest value at SEV and WLS observed in autumn and winter, respectively. The spatial distribution and seasonal evolution of EPE and EKE were consistent with the model results provided in the study by Zhuang et al. (2010).

To understand the differences between EPE and EKE, the energy transfer between these two energy terms and mean

FIG. 10. Seasonal spatial distributions of depth-integrated (a)–(d) T2 ($\text{cm}^3 \text{s}^{-3}$), (e)–(h) EPE ($10^6 \text{cm}^3 \text{s}^{-2}$), (i)–(l) T4 ($\text{cm}^3 \text{s}^{-3}$), and (m)–(p) EKE ($10^6 \text{cm}^3 \text{s}^{-2}$).

potential energy must be examined. Although the vertical transfer term (T3) was not considered, T2 and T4 were examined as an indication of the baroclinic and barotropic instabilities between the energy terms. At SEV, significant positive values of T2 can be observed in all seasons, particularly in

summer and autumn (Figs. 10j,k), and this observation agreed with the high values of EPE and EKE. This finding indicates that the available mean potential energy was converted to EPE through the baroclinic instability, and a part of this EPE was then converted to EKE via T3. In contrast, the most significant

FIG. 11. The baroclinic conversion term (T_2 ; $10^6 \text{ cm}^3 \text{ s}^{-2}$) in the surface layer (0–30 m) during (a) summer and (b) winter. (c),(d) As in (a) and (b), but for the subsurface layer (30–400 m). The lead-lag correlations of the subsurface temperature anomaly (θ' ; contour lines) and subsurface relative vorticity (ξ' ; color shading) along the section [red lines in (c) and (d)] with respect to 117°E during (e) JJA and (f) DJF.

positive values of T_4 were mainly found at WLS and SEV in winter and autumn (Figs. 10o,p). The positive values of T_2 and T_4 were both large at SEV in autumn, implying that both the barotropic and baroclinic instabilities contributed to the increase in EKE.

Significant negative values of T_2 were observed at WLS in all seasons (Figs. 10i–l), with the largest values observed in winter. Further, significant negative values were found at SEV at approximately 12°N, 112°E in spring, autumn, and winter, suggesting an upscale tendency of energy extracted from EPE and returned to the mean flow. This phenomenon not only explains why the EKE at SEV is larger than that at WLS but also indicates an upgradient transport of heat from the perspective of energy, and its location is corresponding to that of the upgradient EHT from the comparison between seasonal mean EHT and temperature gradient (Fig. 7). Significant

negative values of T_4 were also found at the south of SEV, implying that EKE was transferred back to the mean flow.

6. Discussion

a. Upgradient features and possible dynamics of these features at WLS

Usually, the baroclinic instability is a consistent source of the eddy field and is associated with downgradient EHT. However, in this study, an upgradient EHT at WLS is induced by the baroclinic instability through an inverse energy transfer, which is indicated by a negative baroclinic energy conversion term [T_2 , Eq. (9)]. This unique feature is also reported in a recent model study by Xie et al. (2020). Based on climatological forcing model experiment, they attributed this upscale energy

transfer to the intrinsic variability in the ocean. In this section, we aim to discuss the possible dynamics associated with the generation and seasonality of the upgradient EHT at WLS by diagnosing the T2 term.

The first finding is that the largest negative T2 at WLS is located in the subsurface layer, and the associated baroclinic instability at this area is mainly originated from the Kuroshio intrusion. By separating the T2 into surface and subsurface layers based on the relationship between regional mean EHT and temperature gradient (Fig. 8d), it is found that the T2 in the surface layer is positive at most WLS region during summer and winter (Figs. 11a,b), except for a very weak negative T2 area near the Luzon Strait in summer, meaning that the EHT is downgradient in the surface layer. In contrast, the T2 in the subsurface layer shows significant negative values at WLS, indicating an upgradient EHT (Figs. 11c,d). Combining the dynamical analysis of the Kuroshio influence at WLS in section 5c(1), we find that the seasonal variation of T2 in the subsurface layer is associated with the strength of the Kuroshio intrusion. It implies that the largest baroclinic instability at WLS in winter is provided by the Kuroshio intrusion when the zonal velocity shear is the largest and ocean stratification is the weakest (Figs. 9d,h), and vice versa in the summer (Figs. 9b,f). Noted that the Kuroshio also sheds influence on the T2 below the subsurface layer, where it also exhibits similar seasonal variation as that in the subsurface layer, but with at least two orders of magnitude smaller (shading, Fig. S6). Then a natural question next would be how this baroclinic instability induces an inverse energy transfer, which means that the instability is suppressed and absorbed by the mean flow. We find that the largest negative T2 below the subsurface is mainly confined near the Luzon Strait where there is steep topography (red contour, Fig. S6). This result leads to our second finding that the topography may be important in determining the inverse energy transfer, which has been reported from previous studies at the same region (Yang et al. 2013; Xie et al. 2020). Using linear stability analysis or model experiments, several observation and modeling studies at some other regions (de La Lama et al. 2016; LaCasce 2017; Trodahl and Isachsen 2018) also emphasized the role of topography on the baroclinic instability and reached similar conclusion that the steep slope of topography would suppress the vertical scale of instability and inhibit unstable growth of the baroclinic instability.

It is interesting to note that the largest T2 in the subsurface layer at WLS in the winter shows very different spatial structure to those in other seasons. A second subsurface maximum of negative T2 is observed at around 20°N, 115°E (Fig. 11d). Although there is a weak negative T2 below the subsurface at the same area (Fig. S6d), this large negative T2 in the subsurface layer cannot be simply explained by the topography effect as described above because it is far from the source area of baroclinic instability near the Luzon Strait. Using linear stability analysis based on observed climatological temperature and salinity data, Feng et al. (2021) proposed that the subsurface baroclinic instability is associated with propagation of subsurface geostrophic velocity perturbation. Due to the spatial resolution of their data, it is unclear whether or not this propagation of subsurface perturbation exists in the SCS.

Therefore, a lead-lag correlation is performed to the EHT-related variables in the subsurface layer, namely, turbulent temperature (θ') and turbulent relative vorticity (ξ'). To proceed, the lead-lag correlations of subsurface θ' (integrated in the subsurface layer) during winter are calculated at the section where these two largest negative T2 locates (red line, Figs. 11c,d). The subsurface θ' at 117°E is chosen as a reference. Same procedure is also applied to the subsurface ξ' . The lagged correlations show that both the θ' and ξ' (contour line and shading color in Fig. 11e) exhibit a remarkable southwestward-propagating pattern during the winter when these two largest T2 occur. In contrast, both the θ' and ξ' is locally distributed during summer when the T2 is relatively weak and localized. For the wintertime, the typical period of propagation is about 60–70 days, which is consistent with the cospectra period of velocity and temperature anomalies at WLS (Fig. 5b). Moreover, the phase speed is about 10–11 cm s⁻¹, which is comparable to the mean flow velocity in the subsurface layer (8–10 cm s⁻¹) as well as the first mode of Rossby wave. This indicates that the southward propagation is related to the advection of mesoscale eddy by the mean flow or by the Rossby wave, which are reported in previous model and observational studies (Zhuang et al. 2010; Xie et al. 2018; Xie et al. 2020). On the other hand, the winter propagation extends to the Luzon Strait area (120°E) but the correlation is relatively weak, implying that this subsurface propagation is not uniquely originated from the Kuroshio intrusion, and it may be also locally generated by wind stress curl which excites the westward-propagating Rossby wave (Xie et al. 2018), or by the interaction between current and topography (Hu et al. 2008; Wang 2011).

Herein, we can summarize that the baroclinic instability induced upgradient EHT at WLS is mainly generated by the interaction between the Kuroshio intrusion and topography below the surface layer. Particularly, the Kuroshio tends to provide the baroclinic instability, whereas the topography is important in determining the inverse energy transfer. In addition, the most significant upgradient EHT at WLS in winter shows a wave-like southwestward propagation, which is likely caused by the Kuroshio intrusion or the local wind stress and current. However, observation and model with simplified physics are needed to better understand the detailed evolution of instability and the associated perturbations.

b. The rotational and divergent EHT as derived from two different methods

Understanding the difference between EHT_{v θ} and EHT_{dg} is important to achieve some insights into the improvement of eddy parameterization in the SCS. As Marshall and Shutts (1981) pointed out, the rotational part of EHT usually does not contribute to the local heat budget, whereas its divergent part does and is dynamically active. Therefore, it is meaningful to investigate the difference between EHT_{v θ} and EHT_{dg} by separating them into divergent and rotational parts. In this section, we mainly focus on 1) how much EHT consists of these two different components and their relationship with the mean temperature and 2) the role of divergent EHT on the local heat budget.

Equation (12) presents the equations used for this analysis (Jayne and Marotzke 2002). The rotational and divergent EHT

FIG. 12. (a) Rotational and (b) divergent parts of the annual mean depth-integrated $\text{EHT}_{V\theta}$ (vectors) and annual mean depth-integrated temperature (10^4°C ; shading) obtained from the model. (c), (d) As in (a) and (b), but the vectors are EHT_{dg} .

were obtained by first considering the divergence and curl of EHT. Then, two potential functions were recovered by applying an inverse Laplacian operator to the divergence and curl of EHT with Neumann boundary conditions (these conditions assume that there is no heat exchange at the boundaries). Finally, the divergent $(\nabla'\theta')_D$ and rotational $(\nabla'\theta')_R$ parts of EHT were derived by obtaining the gradients of the two potential functions. The rotational and divergent EHT were independent of each other by definition; the divergence (curl) of the rotational (divergent) EHT is zero:

$$\begin{aligned}\nabla'\theta' &= (\nabla'\theta')_R + (\nabla'\theta')_D, \\ (\nabla'\theta')_R &= \hat{k} \times \nabla \nabla^{-2} (\hat{k} \cdot \nabla \times \nabla'\theta'), \\ (\nabla'\theta')_D &= \nabla \nabla^{-2} (\nabla \cdot \nabla'\theta').\end{aligned}\quad (12)$$

In Eq. (12), $\hat{k}$ denotes the vertical direction along the depth, which satisfies the right-handed coordinate.

We applied this method to the regional area with different boundaries to justify the decomposition method. The regionally decomposed rotational and divergent parts of $\text{EHT}_{V\theta}$ (Figs. S5a–d) showed a very similar structure in the same areas as those derived from a larger domain (Fig. 12). The same conclusion can be drawn from the comparison between regionally decomposed EHT_{dg} (Figs. S5e–h) and that from the larger domain. In other words, our decomposition method is relatively robust and can be used to obtain the rotational and divergent parts of EHT.

The values of the divergent and rotational parts of the annual mean $\text{EHT}_{V\theta}$ are presented in Figs. 12a and 12b. In particular, the magnitude of the rotational $\text{EHT}_{V\theta}$ (Fig. 12a) was

significant at SEV, which was expected based on its seasonal mean values (Fig. 7). As proposed by Jayne and Marotzke (2002), the rotational EHT arises from the meandering of the jet stream. Hence, the large rotational $\text{EHT}_{\mathbf{v}_\theta}$ was likely related to the offshore jet stream at SEV, which was driven by the strong wind stress curl in the same area (Wang et al. 2006; Gan and Qu 2008). Conversely, the divergent $\text{EHT}_{\mathbf{v}_\theta}$ was more significant at WLS, with a widespread pattern (Fig. 12b). The rotational and divergent parts of $\text{EHT}_{\mathbf{v}_\theta}$ are then compared with those of EHT_{dg} (Figs. 12c,d). Note that before the analysis, compared with the EHT_{dg} estimated by Chen et al. (2012), our findings produced an underestimate in the EHT magnitude because in this case, EHT_{dg} included only short-term perturbations (Fig. S3). A large contrast between the rotational parts of $\text{EHT}_{\mathbf{v}_\theta}$ and EHT_{dg} can be observed. One possible reason for this significant difference can be associated with the “divergent nature” of EHT_{dg} (Jayne and Marotzke 2002), which can be proved by taking the curl of $-\kappa \nabla_h \theta$. The resultant two terms were $\kappa \nabla \times \nabla \theta$ and $\nabla \kappa \times \nabla \theta$, respectively. The first term equals zero, and the second term has a very small value because both $\nabla \kappa$ and $\nabla \theta$ tend to follow the baroclinic fronts (Stammer 1998).

By combining two components of EHT with depth-integrated temperature (vectors and color shading, Fig. 12), it is found that both the rotational and divergent $\text{EHT}_{\mathbf{v}_\theta}$ tend to flow in the upgradient or downgradient directions. In particular, the divergent $\text{EHT}_{\mathbf{v}_\theta}$ exhibits a more consistent downgradient feature at SEV, meaning that it flows in the opposite direction with that of temperature gradient, which directs southeastward normal to the isotherms. Conversely, the divergent $\text{EHT}_{\mathbf{v}_\theta}$ shows a more consistent upgradient feature at the western boundary area near WLS, where the EHT flows in the same direction with that of temperature gradient, which directs northwestward normal to the isotherms. In contrast, the rotational $\text{EHT}_{\mathbf{v}_\theta}$ does not exhibit an obvious upgradient feature at WLS, implying that the divergent part of $\text{EHT}_{\mathbf{v}_\theta}$ is more important in determining this upgradient EHT.

The divergent parts of both $\text{EHT}_{\mathbf{v}_\theta}$ and EHT_{dg} showed a significant difference at WLS. However, their divergent parts share a similar spatial structure in the surface mixed layer, where they are related to the surface heat flux (Fig. S7). More importantly, even though the divergent total heat transport (Fig. S7e) exhibits much larger magnitude than the divergent EHT, the divergence of $\text{EHT}_{\mathbf{v}_\theta}$ is more spatially correlated with the positive turbulent heat flux (Fig. S7b), indicating that the heat loss at the sea surface is mainly related to the divergence of $\text{EHT}_{\mathbf{v}_\theta}$. Meanwhile, the large convergence of $\text{EHT}_{\mathbf{v}_\theta}$ at SEV is associated with the negative radiative heat flux at the same place, meaning that heat gain by the ocean at SEV is mainly associated with the convergence of $\text{EHT}_{\mathbf{v}_\theta}$ (Fig. S7a). It is shown that the divergent $\text{EHT}_{\mathbf{v}_\theta}$ exhibits an upgradient feature as a depth integral but exhibits downgradient feature in the surface layer, implying that this upgradient EHT may be associated with the divergent EHT in the subsurface layer.

7. Conclusions

The EHT in SCS was investigated using an eddy-resolving assimilated model. The assimilated model output results agreed

with the daily temperature profiles obtained from the Argo floats as well as with the daily velocity profiles obtained using the three mooring buoys. Moreover, the model could reproduce the EHT_{dg} -related fields from observations, which corresponded to the surface EKE and depth-integrated temperature gradient. In contrast to previous studies, the time-varying $\text{EHT}_{\mathbf{v}_\theta}$ used herein had a period of less than 100 days.

The most significant EKE was found at SEV, where the covariability of the temperature and velocity anomalies had a period of 70 days. A second significant EKE area was located at WLS, which had a covariability with a period of 55 days. In particular, the depth-integrated annual mean EHT showed significant northwestward EHT at both SEV and WLS. By separating the EHT into different layers, we observed that most EHT was confined to the top 1000 m, with the largest EHT in the top 400 m. Furthermore, a significant seasonality was observed in the depth-integrated EHT, with the largest value at SEV and WLS in autumn and winter, respectively. The strong seasonality in EHT in these two regions was mainly observed between 50 and 200 m, and the largest amounts of EHT were found at shallower depths at SEV than at WLS. Additionally, an upgradient EHT, which indicated that the EHT and temperature gradient were in the same direction, was identified at WLS, mainly below 50 m, particularly for the annual and seasonal means of MEHT.

The seasonal variation in EHT at SEV was related to the local wind stress curl, whereas that at WLS was associated with the Kuroshio intrusion, which was represented by the buoyancy frequency and zonal velocity shear. An energy budget analysis revealed that both the baroclinic and barotropic instabilities were important for generating EKE at SEV in autumn. Meanwhile, the barotropic instability was mainly responsible for the increase in EKE at WLS in winter. The upgradient EHT at WLS was further confirmed by the baroclinic instability through an inverse energy transfer, which was generated by the interaction between the Kuroshio intrusion and topography below the surface layer. Moreover, the most significant upgradient EHT in winter showed a wave-like southwestward-propagating pattern in the subsurface layer.

Finally, by separating EHT into rotational and divergent parts, the difference between $\text{EHT}_{\mathbf{v}_\theta}$ and EHT_{dg} was investigated. Two conclusions can be drawn from the comparison between the rotational and divergent parts of $\text{EHT}_{\mathbf{v}_\theta}$ and EHT_{dg} : 1) the rotational part of $\text{EHT}_{\mathbf{v}_\theta}$, which EHT_{dg} lacked, was also important to the actual $\text{EHT}_{\mathbf{v}_\theta}$, and 2) the large difference between divergent $\text{EHT}_{\mathbf{v}_\theta}$ and EHT_{dg} was mainly observed at the subsurface layer. However, they shared a similar spatial pattern in the mixed layer, where they were related to the air–sea heat flux. These two findings should be considered to improve the parameterization of EHT.

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REFERENCES

- Adcroft, A., and Coauthors, 2018: MITgcm user manual. Massachusetts Institute of Technology, 485 pp., http://mitgcm.org/public/r2_manual/final/online_documents/manual.html.
- Atlas, R., R. N. Hoffman, J. Ardizzone, S. M. Leidner, J. C. Jusem, D. K. Smith, and D. Gombos, 2011: A cross-calibrated, multiplatform ocean surface wind velocity product for meteorological and oceanographic applications. *Bull. Amer. Meteor. Soc.*, **92**, 157–174, <https://doi.org/10.1175/2010BAMS2946.1>.
- Balmaseda, M. A., K. Mogensen, and A. T. Weaver, 2013: Evaluation of the ECMWF ocean reanalysis system ORAS4. *Quart. J. Roy. Meteor. Soc.*, **139**, 1132–1161, <https://doi.org/10.1002/qj.2063>.
- Barnier, B., L. Siefridt, and P. Marchesiello, 1995: Thermal forcing for a global ocean circulation model using a three-year climatology of ECMWF analyses. *J. Mar. Syst.*, **6**, 363–380, [https://doi.org/10.1016/0924-7963\(94\)00034-9](https://doi.org/10.1016/0924-7963(94)00034-9).
- Beckmann, A., C. W. Böning, B. Brügge, and D. Stammer, 1994: On the generation and role of eddy variability in the central North Atlantic Ocean. *J. Geophys. Res.*, **99**, 20 381–20 391, <https://doi.org/10.1029/94JC01654>.
- Bleck, R., 2002: An oceanic general circulation model framed in hybrid isopycnic-Cartesian coordinates. *Ocean Modell.*, **4**, 55–88, [https://doi.org/10.1016/S1463-5003\(01\)00012-9](https://doi.org/10.1016/S1463-5003(01)00012-9).
- Böning, C. W., and M. D. Cox, 1988: Particle dispersion and mixing of conservative properties in an eddy-resolving model. *J. Phys. Oceanogr.*, **18**, 320–338, [https://doi.org/10.1175/1520-0485\(1988\)018<0320:PDAMOC>2.0.CO;2](https://doi.org/10.1175/1520-0485(1988)018<0320:PDAMOC>2.0.CO;2).
- , and R. G. Budich, 1992: Eddy dynamics in a primitive equation model: Sensitivity to horizontal resolution and friction. *J. Phys. Oceanogr.*, **22**, 361–381, [https://doi.org/10.1175/1520-0485\(1992\)022<0361:EDIAPE>2.0.CO;2](https://doi.org/10.1175/1520-0485(1992)022<0361:EDIAPE>2.0.CO;2).
- Bryden, H. L., 1979: Poleward heat flux and conversion of available potential energy in Drake Passage. *J. Mar. Res.*, **37**, 1–22.
- Casey, K. S., T. B. Brandon, P. Cornillon, and R. Evans, 2010: The past, present, and future of the AVHRR Pathfinder SST Program. *Oceanography from Space*, V. Barale, J. F. R. Gower, and L. Alberotanza, Eds., Springer, 273–287, https://doi.org/10.1007/978-90-481-8681-5_16.
- Chakraborty, A., R. Kumar, S. Basu, and R. Sharma, 2015: Improving ocean state by assimilating SARAL/AltiKa derived sea level and other satellite-derived data in MITGCM. *Mar. Geod.*, **38**, 328–338, <https://doi.org/10.1080/01490419.2014.1002142>.
- Chassignet, E. P., and Coauthors, 2007: The HYCOM (Hybrid Coordinate Ocean Model) data assimilative system. *J. Mar. Syst.*, **65**, 60–83, <https://doi.org/10.1016/j.jmarsys.2005.09.016>.
- Chelton, D. B., M. G. Schlax, R. M. Samelson, and R. A. de Szoeke, 2007: Global observations of large oceanic eddies. *Geophys. Res. Lett.*, **34**, L15606, <https://doi.org/10.1029/2007GL030812>.
- Chen, G., Y. Hou, Q. Zhang, and X. Chu, 2010: The eddy pair off eastern Vietnam: Interannual variability and impact on thermohaline structure. *Cont. Shelf Res.*, **30**, 715–723, <https://doi.org/10.1016/j.csr.2009.11.013>.
- , —, and —, 2011: Mesoscale eddies in the South China Sea: Mean properties, spatiotemporal variability, and impact on thermohaline structure. *J. Geophys. Res.*, **116**, C06018, <https://doi.org/10.1029/2010JC006716>.
- , J. Gan, Q. Xie, X. Chu, D. Wang, and Y. Hou, 2012: Eddy heat and salt transports in the South China Sea and their seasonal modulations. *J. Geophys. Res.*, **117**, C05021, <https://doi.org/10.1029/2011JC007724>.
- Chu, P. C., C. W. Fan, C. J. Lozano, and J. L. Kerling, 1998: An airborne expendable bathythermograph survey of the South China Sea, May 1995. *J. Geophys. Res.*, **103**, 21 637–21 652, <https://doi.org/10.1029/98JC02096>.
- Cooper, M., and K. Haines, 1996: Altimetric assimilation with water property conservation. *J. Geophys. Res.*, **101**, 1059–1077, <https://doi.org/10.1029/95JC02902>.
- de La Loma, M., J. H. LaCasce, and H. K. Fuhr, 2016: The vertical structure of ocean eddies. *Dyn. Stat. Climate Syst.*, **1**, dzw001, <https://doi.org/10.1093/climsys/dzw001>.
- Durski, S. M., S. M. Glenn, and D. B. Haidvogel, 2004: Vertical mixing schemes in the coastal ocean: Comparison of the level 2.5 Mellor–Yamada scheme with an enhanced version of the K profile parameterization. *J. Geophys. Res.*, **109**, C01015, <https://doi.org/10.1029/2002JC001702>.
- Feng, L., C. Liu, A. Köhl, D. Stammer, and F. Wang, 2021: Four types of baroclinic instability waves in the global oceans and the implications for the vertical structure of mesoscale eddies. *J. Geophys. Res. Oceans*, **126**, e2020JC016966, <https://doi.org/10.1029/2020JC016966>.
- Gan, J., and T. Qu, 2008: Coastal jet separation and associated flow variability in the southwest South China Sea. *Deep-Sea Res. I*, **55**, 1–19, <https://doi.org/10.1016/j.dsr.2007.09.008>.
- Hsin, Y. C., C. R. Wu, and P. T. Shaw, 2008: Spatial and temporal variations of the Kuroshio east of Taiwan, 1982–2005: A numerical study. *J. Geophys. Res.*, **113**, C04002, <https://doi.org/10.1029/2007JC004485>.
- Hu, X., X. Xiong, F. Qiao, B. Guo, and X. Lin, 2008: Surface current field and seasonal variability in the Kuroshio and adjacent regions derived from satellite-tracked drifter data. *Acta Oceanol. Sin.*, **27** (3), 11–29.

- Jayne, S. R., and J. Marotzke, 2002: The oceanic eddy heat transport. *J. Phys. Oceanogr.*, **32**, 3328–3345, [https://doi.org/10.1175/1520-0485\(2002\)032<3328:TOEHT>2.0.CO;2](https://doi.org/10.1175/1520-0485(2002)032<3328:TOEHT>2.0.CO;2).
- Jiang, W., Z. Cheng, B. Xing, and J. Wang, 2016: Characteristic analysis of spatiotemporal variability of eddy heat transports in the South China Sea (in Chinese with English abstract). *Hydrogr. Surv. Charting*, **36**, 35–38.
- Kalnay, E., and Coauthors, 1996: The NCEP/NCAR 40-Year Reanalysis Project. *Bull. Amer. Meteor. Soc.*, **77**, 437–471, [https://doi.org/10.1175/1520-0477\(1996\)077<0437:TNYRP>2.0.CO;2](https://doi.org/10.1175/1520-0477(1996)077<0437:TNYRP>2.0.CO;2).
- Kuhlbrodt, T., and J. M. Gregory, 2012: Ocean heat uptake and its consequences for the magnitude of sea level and climate change. *Geophys. Res. Lett.*, **39**, L18608, <https://doi.org/10.1029/2012GL052952>.
- LaCasce, J. H., 2017: The prevalence of oceanic surface modes. *Geophys. Res. Lett.*, **44**, 11 097–11 105, <https://doi.org/10.1002/2017GL075430>.
- Large, W. G., J. C. McWilliams, and S. C. Doney, 1994: Oceanic vertical mixing: A review and a model with nonlocal boundary layer parameterization. *Rev. Geophys.*, **32**, 363–403, <https://doi.org/10.1029/94RG01872>.
- Liang, W.-D., Y. J. Yang, T. Y. Tang, and W.-S. Chuang, 2008: Kuroshio in the Luzon Strait. *J. Geophys. Res.*, **113**, C08048, <https://doi.org/10.1029/2007JC004609>.
- Locarnini, R. A., and Coauthors, 2013: *Temperature*. Vol. 1, *World Ocean Atlas 2013*, NOAA Atlas NESDIS 73, 40 pp., https://data.nodc.noaa.gov/wao/WOA13/DOC/wao13_voll.pdf.
- Lorenz, E. N., 1955: Available potential energy and the maintenance of the general circulation. *Tellus*, **7**, 157–167, <https://doi.org/10.3402/tellusa.v7i2.8796>.
- Luo S. H., Z. Y. Jing, Y. Q. Qi, and Q. Xie, 2016: Numerical study on sub-mesoscale processes in the northern South China Sea (in Chinese with English abstract). *J. Trop. Oceanogr.*, **35** (5), 10–19.
- Macdonald, A. M., and C. Wunsch, 1996: An estimate of global ocean circulation and heat fluxes. *Nature*, **382**, 436–439, <https://doi.org/10.1038/382436a0>.
- Marshall, J., and G. Shutts, 1981: A note on rotational and divergent eddy fluxes. *J. Phys. Oceanogr.*, **11**, 1677–1680, [https://doi.org/10.1175/1520-0485\(1981\)011<1677:ANORAD>2.0.CO;2](https://doi.org/10.1175/1520-0485(1981)011<1677:ANORAD>2.0.CO;2).
- , A. Adcroft, C. Hill, L. Perelman, and C. Heisey, 1997: A finite-volume, incompressible Navier–Stokes model for studies of the ocean on parallel computers. *J. Geophys. Res.*, **102**, 5753–5766, <https://doi.org/10.1029/96JC02775>.
- NOAA/NGDC, 2006: 2-minute Gridded Global Relief Data (ETOPO2) v2. NOAA/National Geophysical Data Center, accessed June 2006, <https://doi.org/10.7289/V5J1012Q>.
- Pan, J. Y., and Y. J. Sun, 2018: Estimation of horizontal eddy heat flux in upper mixed-layer in the South China Sea by using satellite data. *Sci. Rep.*, **8**, 15527, <https://doi.org/10.1038/s41598-018-33803-2>.
- Qiu, B., and S. Chen, 2005: Eddy-induced heat transport in the subtropical North Pacific from Argo, TMI, and altimetry measurements. *J. Phys. Oceanogr.*, **35**, 458–473, <https://doi.org/10.1175/JPO2696.1>.
- Roemmich, D., J. Gilson, B. Cornuelle, and R. Weller, 2001: Mean and time-varying meridional transport of heat at the tropical/subtropical boundary of the North Pacific Ocean. *J. Geophys. Res.*, **106**, 8957–8970, <https://doi.org/10.1029/1999JC000150>.
- Rossby, T., 1987: On the energetics of the Gulf Stream at 73W. *J. Mar. Res.*, **45**, 59–82, <https://doi.org/10.1357/002224087788400918>.
- Sheu, W. J., C. R. Wu, and L. Y. Oey, 2010: Blocking and westward passage of eddies in the Luzon Strait. *Deep-Sea Res. II*, **57**, 1783–1791, <https://doi.org/10.1016/J.DSR2.2010.04.004>.
- Smagorinsky, J., 1963: General circulation experiments with the primitive equations: I. The basic experiment. *Mon. Wea. Rev.*, **91**, 99–164, [https://doi.org/10.1175/1520-0493\(1963\)091<0099:GCEWTP>2.3.CO;2](https://doi.org/10.1175/1520-0493(1963)091<0099:GCEWTP>2.3.CO;2).
- Stammer, D., 1998: On eddy characteristics, eddy transports, and mean flow properties. *J. Phys. Oceanogr.*, **28**, 727–739, [https://doi.org/10.1175/1520-0485\(1998\)028<0727:OECETA>2.0.CO;2](https://doi.org/10.1175/1520-0485(1998)028<0727:OECETA>2.0.CO;2).
- Su, J., 2005: Overview of the South China Sea circulation and its dynamics. *Acta Oceanol. Sin.*, **27** (6), 1–8.
- Sun, B., C. Liu, and F. Wang, 2019: Global meridional eddy heat transport inferred from Argo and altimetry observations. *Sci. Rep.*, **9**, 1345, <https://doi.org/10.1038/s41598-018-38069-2>.
- Tang, Y., W. W. Hsieh, B. Tang, and K. Haines, 2001: A neural network atmospheric model for hybrid coupled modelling. *Climate Dyn.*, **17**, 445–455, <https://doi.org/10.1007/s003820000119>.
- Trodahl, M., and P. E. Isachsen, 2018: Topographic influence on baroclinic instability and the mesoscale eddy field in the northern North Atlantic Ocean and the Nordic Seas. *J. Phys. Oceanogr.*, **48**, 2593–2607, <https://doi.org/10.1175/JPO-D-17-0220.1>.
- Volkov, D. L., T. Lee, and L. L. Fu, 2008: Eddy-induced meridional heat transport in the ocean. *Geophys. Res. Lett.*, **35**, L20601, <https://doi.org/10.1029/2008GL035490>.
- Wang, D. K., 2011: Numerical study on eddy kinetic energy and eddy heat transport in the South China Sea (in Chinese with English abstract). Ph.D. thesis, Ocean University of China, 119 pp.
- Wang, G., J. Su, and P. C. Chu, 2003: Mesoscale eddies in the South China Sea observed with altimeter data. *Geophys. Res. Lett.*, **30**, 2121, <https://doi.org/10.1029/2003GL018532>.
- , —, and Y. Q. Quan, 2005: Advances in studying mesoscale eddies in South China Sea (in Chinese with English abstract). *Adv. Earth Sci.*, **20**, 882–886.
- , D. Chen, and J. Su, 2006: Generation and life cycle of the dipole in the South China Sea summer circulation. *J. Geophys. Res.*, **111**, C06002, <https://doi.org/10.1029/2005JC003314>.
- Wu, C. R., and T. L. Chiang, 2007: Mesoscale eddies in the northern South China Sea. *Deep-Sea Res. II*, **54**, 1575–1588, <https://doi.org/10.1016/j.dsr2.2007.05.008>.
- Wunsch, C., 1999: Where do ocean eddy heat fluxes matter? *J. Geophys. Res.*, **104**, 13 235–13 249, <https://doi.org/10.1029/1999JC900062>.
- Xie, J., L. Zhou, R. Murtugudde, M. Jochum, and Y. Zhong, 2020: Internal subseasonal variability in the South China Sea revealed by eddy-resolving numerical simulations. *J. Geophys. Res. Oceans*, **125**, e2019JC015390, <https://doi.org/10.1029/2019JC015390>.
- Xie, L., Q. Zheng, S. Zhang, J. Hu, M. Li, J. Li, and Y. Xu, 2018: The Rossby normal modes in the South China Sea deep basin evidenced by satellite altimetry. *Int. J. Remote Sens.*, **39**, 399–417, <https://doi.org/10.1080/01431161.2017.1384591>.
- Yang, H., L. Wu, H. Liu, and Y. Yu, 2013: Eddy energy sources and sinks in the South China Sea. *J. Geophys. Res. Oceans*, **118**, 4716–4726, <https://doi.org/10.1002/jgrc.20343>.
- Yang, Y., and Coauthors, 2019: Eddy-induced transport of saline Kuroshio water into the northern South China Sea. *J. Geophys. Res. Oceans*, **124**, 6673–6687, <https://doi.org/10.1029/2018JC014847>.

- Yim, B. Y., and Coauthors, 2010: The Vertical structure of eddy heat transport simulated by an eddy-resolving OGCM. *J. Phys. Oceanogr.*, **40**, 340–353, <https://doi.org/10.1175/2009JPO4243.1>.
- Yuan, D., W. Han, and D. Hu, 2006: Surface Kuroshio path in the Luzon Strait area derived from satellite remote sensing data. *J. Geophys. Res.*, **111**, C11007, <https://doi.org/10.1029/2005JC003412>.
- Zhang, Z., W. Zhao, B. Qiu, and J. Tian, 2017: Anticyclonic eddy sheddings from Kuroshio loop and the accompanying cyclonic eddy in the northeastern South China Sea. *J. Phys. Oceanogr.*, **47**, 1243–1259, <https://doi.org/10.1175/JPO-D-16-0185.1>.
- Zhuang, W., S. P. Xie, D. Wang, B. Taguchi, H. Aiki, and H. Sasaki, 2010: Intraseasonal variability in sea surface height over the South China Sea. *J. Geophys. Res.*, **115**, C04010, <https://doi.org/10.1029/2009JC005647>.