Impacts of Wind and Current on the Interannual Variation of the Summertime Upwelling Off Southern Vietnam in the South China Sea

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Abstract Summertime upwelling off the southern Vietnamese coast is one of the most essential oceanographic features in the South China Sea. Based on analyzing the 38-year (1982–2019) sea surface temperature (SST) images, locations of summertime Vietnamese upwelling centers are found to be classified into three sub-regions: the Southern Coastal Upwelling (SCU; south of 12.5°N), the Northern Coastal Upwelling (NCU; north of 12.5°N), and the Offshore Upwelling (OU; east of 110°E). Variations of upwelling intensities in the three sub-regions are further quantified via an adaptive SST-based upwelling index, and possible processes relevant to wind field (including wind stress and its curl) and currents are, respectively, proposed. The analyses show that the local along-shore wind stress, inducing offshore Ekman transport, can produce the coastal upwelling-favorable condition but is not the only one of main factors in controlling the interannual variability of the coastal upwelling. The enhancement of wind stress curl dipole off the southern Vietnamese coast, which accompanies the scaling up of the double-gyre structure east of Vietnam and results in the reinforcement and southward shift of eastward-flowing jet, and the exaggeration of the cyclonic gyre-associated southward along-shore current are responsible for the intensification of upwelling in the SCU, but suppress the development of upwelling in the NCU. The well-developed double-gyre structure serves as the essential condition for the OU occurrence and the OU is much more sensitive to the change of cyclonic gyre to the north of the jet.

Plain Language Summary Upwelling is one of the most important oceanographic features, and it plays important roles in regulating local and global climates and marine ecosystem. In general, past research has suggested that upwelling off eastern Vietnam in the South China Sea occurs in summer and significantly suffers interannual changes, and wind forcing, along-shore current, and eastward-flowing jet off the eastern Vietnam could be the controlling factors. In this study, we explore the Vietnamese upwelling in detail by separating the entire upwelling area into three sub-regions (the southern Vietnameses coast, the northern Vietnameses coast, and the offshore region) and clarify the roles of controlling factors in each region. Along the Vietnamese coastal regions, the local along-shore wind stress can produce the coastal upwelling-favorable condition but is not the main factor. The enhancement of wind stress curl dipole off the southern Vietnamese coast and the southward along-shore current north of ~12°N are responsible for the intensification of upwelling in the southern coast, but suppress the development of upwelling in the northern coast. In the offshore region, the well-developed double of oceanic gyre is the essential condition and the change of cyclonic gyre to the north of the jet plays a more important role.

1. Introduction

The South China Sea (SCS) is the largest marginal sea of the western Pacific Ocean. Adjoining the western Pacific Ocean and the eastern Indian Ocean, the atmospheric and the oceanic dynamical processes of the SCS are strongly influenced by climate variability of the two oceans (Liu et al., 2014; Palacz et al., 2017; Qu et al., 2009; Yang et al., 2015). The wind regime in the SCS is controlled by the East Asian monsoon with northeasterly wind in winter, and southwesterly wind in summer (Xie et al., 2003). Driven by the monsoon, the surface circulation in the SCS is also characterized by seasonally alternating basin-scale gyres, basically a cyclonic gyre in winter and a cyclonic northern gyre and an anticyclonic southern gyre with an
eastward/northeastward jet in-between and shooting from the Vietnam coast in summer (Fang, 2002; Gan & Qu, 2008; Gan et al., 2006).

Upwelling is one of the most prominent oceanographic features in the SCS. The upwelling in the SCS can modulate local atmospheric condition by causing oceanic feedback into the local atmosphere (e.g., reducing the local wind speed) (Sun et al., 2016; Xie et al., 2007; Zheng et al., 2016), is responsible for the renewal of the deep water (Chao et al., 1996a, 1996b; Zhu et al., 2016) and the enhancement of ecological activities (Hein et al., 2013; Isoguchi & Kawamura, 2006; Liu et al., 2012; Voss et al., 2006; Xie et al., 2003; Zhao & Tang, 2007). Besides, the SCS upwelling highly couples with the equatorial Pacific variability (Xie et al., 2003). There are two typical upwelling regions identified in the SCS; one is off the northwestern Luzon Island and the other is located east of Vietnam (Chao et al., 1996b; Shaw et al., 1996; Tan et al., 2009; Wyrtki, 1961). The former upwelling occurs in winter (October-January) and is proposed to be driven by offshore Ekman drift in the surface layer and a northward undercurrent converging in the subsurface layer forced by basin-scale circulation (Chao et al., 1996b; Shaw et al., 1996; Chen et al., 2006).

The upwelling off eastern Vietnam, which is the focus of this study, develops in May–September and reaches the mature phase in July–August (Fang et al., 2012; Liu et al., 2002; Voss et al., 2006). Past studies suggested that this summertime upwelling always appears as a jet-shape cold tongue (or cold patch), which originates between 9°N and 15°N along the Vietnamese coast (Dippner et al., 2007; Kuo et al., 2000; Hein et al., 2013; Xie et al., 2003). After generating, the cold tongue could stretch eastward or northeastward into the central SCS (Gan et al., 2006; Xie et al., 2003). Therefore, cold patches associated with the summertime upwelling have been found in both the Vietnamese coastal region and the offshore region east of Vietnam (Chao et al., 1996a; Dippner & Loick-Wilde, 2011; Hu & Wang, 2016; Lin et al., 2017; Xie et al., 2003). In the offshore region, the upwelling usually accompanies the appearance of a cold (or cyclonic) eddy, which is known as the cyclonic East Vietnam Eddy or the Vietnam Cold Eddy (Hu & Wang, 2016; Liu et al., 2008; Xie et al., 2003; Xiu & Chai, 2011). In addition to cold waters observed in the upwelling regions, low sea surface height and high chlorophyll-a concentration were frequently reported (Ho et al., 2000a; Li et al., 2014; Shaw & Chao, 1994; Tan & Shi, 2009; Tang et al., 2004; Xie et al., 2003; Zhao & Tang, 2007).

Although the summertime Vietnamese upwelling is regarded as a seasonal phenomenon, its evolution experiences significant intraseasonal and interannual variations as well. Xie et al. (2007) proposed that the Vietnamese upwelling undergoes 2–4 intraseasonal cycles of development and decay. The intraseasonal change is ascribed to the intensification and relaxation of wind pulse in a period of 2 weeks (Xie et al., 2007; Mao & Wang, 2018).

On the interannual timescale, the focused scale of this study, the summertime Vietnamese upwelling has been suggested to be related to the El Niño and Southern Oscillation (ENSO) (Dippner et al., 2007; Hein et al., 2013; Kuo et al., 2004; Liu et al., 2012; Voss et al., 2006; Wu et al., 1998; Xie et al., 2003; Zhao & Tang, 2007). By correlating the June–August cold “filament” index and January Nino3 index, Xie et al. (2003) found the ENSO leads the summertime Vietnamese cold tongue by a half-year period. Kuo et al. (2004) indicated that the southwesterly winds intensified the upwelling and advected the upwelling center southward during the 1997 El Niño, but weakened the upwelling during the following 1998 La Niña. Hein et al. (2013) claimed that, during summer after an El Niño event, the surface cooling triggered by the Vietnamese upwelling tends to be less effective since the upwelled waters originate from shallower depths.

In sum, the Vietnamese upwelling becomes weaker in the summer following an El Niño event in the equatorial Pacific Ocean maturing in winter. Two major factors have been proposed to govern the summertime Vietnamese upwelling on a variety of timescales; one is the southwesterly wind forcing and the other is the eastward/northeastward-flowing jet associated with the dipole gyre east of the central Vietnamese coast (Chen et al., 2012; Dippner et al., 2007; Fang et al., 2012; Hein, 2008; Ho et al., 2000a, 2000b; Hu & Wang, 2016; Kuo et al., 2000; Li et al., 2014; Metzger, 2003; Voss et al., 2006; Xie et al., 2003; Wyrtki, 1961). The wind forcing may exert the impact on the upwelling in the forms of wind stress (i.e., Ekman transport) or wind stress curl (i.e., Ekman pumping). The coastal upwelling is primarily driven by cross-shore Ekman transport induced by the alongshore component of the summer southwesterly monsoon (Chen et al., 2012; Kuo et al., 2000; Metzger, 2003; Tang et al., 2004). Whereas, the open ocean wind-induced Ekman pumping generated by positive wind stress curl and the basin-scale circulation could control the upwelling in
the offshore area (Chu et al., 1998; Hein et al., 2013; Lin et al., 2017; Liu et al., 2008; Liu et al., 2012; Xie et al., 2003; Zhao & Tang, 2006). Regarding the eastward-flowing jet which separates from the Vietnamese coast at about 11°N, previous literature has pointed out the vital role of this jet in advecting the cold coastal water offshore (Bayler & Liu, 2008; Kuo et al., 2000; Liu et al., 2008; Xie et al., 2007). Furthermore, the coincidence of the upward movement of water with the position of the eastward-flowing jet leaving the Vietnamese coast indicated that the separation of the jet could induce the coastal upwelling as well (Chao et al., 1996a; Lin et al., 2017; Liu et al., 2002; Liu et al., 2010).

To sum up, the interannual change of the upwelling has been related to the occurrence of ENSO, and the associated weakening of basin-scale wind is thought of as the major regulator. Although the Vietnamese summertime upwelling has been extensively investigated in the literature, it is always discussed in a synoptic view and detail examinations are still limited. Thus, the study first aims to revisit the interannual variability of summertime Vietnamese upwelling by analyzing an adapted upwelling index derived from remote-sensed sea surface temperature (SSTUI; see Section 3 for details) in the period from 1982 to 2019 and to discuss the underlying dynamical processes in detail by analyzing the dynamical upwelling indices or factors. In the Vietnamese summertime upwelling region, appearance frequency and intensity of upwelling vary significantly depending on location. Therefore, we classified the whole upwelling region into three sub-regions. In this regard, the other goal in the study is to clarify the relative importance of wind forcing (including wind stress and wind stress curl) and the summertime eastward-flowing jet off the southern Vietnamese coast to the interannual changes of occurrences and intensities of the Vietnamese upwelling in the three sub-regions.

The study is structured as follows. Descriptions of data and quantities adopted in this study are briefly introduced in Sections 2 and 3, respectively. Section 4 presents the interannual variations of the upwelling activities in the coastal and offshore regions east of Vietnam, and further clarifies the factors controlling the interannual change of upwelling intensity and discusses the underlying physical processes for each sub-region. The final section summarizes the main findings of this study.

2. Data

In this study, we use remote-sensed sea surface temperature (SST) images to offer a comprehensive view of the interannual changes of upwelling activities off the Vietnamese coast. We further adopt sea surface wind velocity and satellite-altimeter-derived absolute dynamic topography (ADT) to characterize the contribution of governing factors (including wind and ocean current) to the upwelling variability. All the above-mentioned satellite data is temporally averaged over the months between June and September from 1982 to 2019 and, hereafter, the period of June–September is referred to as summertime.

2.1. Sea Surface Temperature

The SST data is based on the emission of radiations in the infrared and microwave bands, which come from depths of ≈10 µm and ≈1 mm, respectively, below the sea surface. The infrared radiometer provides high resolution and high accuracy SST images (Castro et al., 2004) but has its limitations as a result of the presence of clouds, water vapor and atmospheric aerosols (Gentemann et al., 2004). In contrast, microwave measurement is insensitive to aerosols and not obstructed by clouds. It is, however, affected by high-speed winds and precipitation, and cannot detect oceanic signals near coastal areas due to contamination of signals from the land (Gentemann et al., 2004; Guan & Kawamura, 2003). Therefore, we mainly use the infrared SST data because of its longer availability (38 years, 1982–2019), and perform a cross-validation with the microwave SST to obtain a more reliable understanding of SST distribution.

The daily 0.25 × 0.25 infrared SST data in 1982–2019 is downloaded from the Optimum Interpolation SST Version 2 High Resolution Dataset (OISST V2), distributed by the National Oceanic and Atmospheric Administration Earth System Research Laboratory’s Physical Sciences Division (NOAA ESRL PSD; https://www.esrl.noaa.gov/psd/data/gridded/data.noaa.oisst.v2.highres.html). The microwave SST images are obtained from the Group for High Resolution Sea Surface Temperature (GHRSSST) which merged multi-microwave sensors (TMI, AMSRE, GMI, AMSR2, and WindSat) by utilizing the optimal interpolation method. The GHRSSST data is produced daily by the Remote Sensing Systems (REMS; ftp://remss.com/SST/daily/mw/v05.0/netcdf/)

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at a spatial resolution of 0.25° after 1998. In the shallow waters and the vicinity of islands, the GHRSST data is masked as missing value.

In a quantitative manner of validation, the correlation between spatial patterns of the summertime NOAA OISST and GHRSST over the SCS (100°E–120°E and 3°N–22°N) in the overlapping period of 1998–2019 is calculated. The yearly pattern correlation coefficient between the two SST datasets ranges between 0.7 and 0.97 with a mean of 0.89, which is significant at the 99% confidence level of the student’s t-test (degree of freedom = 3213).

Similarly, the spatial pattern correlation analysis is also applied to the Vietnamese upwelling-affected region (105.5°E–116.5°E and 8.5°N–17°N) to further confirm the validity of the SST in our focused area. The yearly pattern correlation coefficient has a range between 0.48 and 0.97 and its mean is 0.82, which is significant at the 99% confidence level of the student’s t-test (degree of freedom = 760). In general, the above quantitative comparison indicates the two datasets can present a robust year-to-year change of summertime SST in the SCS. Therefore, we can utilize the summertime SST to quantify the upwelling intensity (introduced in Section 3.1).

2.2. Wind

Monthly 0.25 × 0.25 wind data at the 10-m height above the sea surface (10-m wind) is obtained from a global atmosphere reanalysis product, the ERA5, which is provided by the European Center for Medium-Range Weather Forecasts (ECMWF; https://www.ecmwf.int/en/forecasts/datasets/reanalysis-datasets/era5/) from 1979 to 2019. The 10-m wind \((U_{10}, V_{10})\) is then converted to wind stress based on the bulk formula of

\[
\tau_x, \tau_y = \rho_d C_d (U_{10}, V_{10}) \sqrt{U_{10}^2 + V_{10}^2}
\]

with nonlinear drag coefficients \((C_d)\) suggested by Trenberth et al. (1990), air density \((\rho_d)\) of 1.3 kg m\(^{-3}\), and x and y are longitude and latitude. Wind stress on land is set as missing value, and wind stress curl \((\nabla \times \boldsymbol{\tau})\) is then calculated based on the central difference method. Afterward, the wind stress and its curl are adopted for discussing the dynamical processes involving in the interannual change of summertime upwelling off the southern Vietnamese coast.

2.3. ADT and Surface Geostrophic Current

ADT is the sea surface height above geoid and calculated by summing up the altimetry sea level anomaly and mean oceanic dynamic topography. The ADT data used in this study is obtained from http://www.aviso. altimetry.fr/en/home.html. It is produced by the Segment Sol multimission d’ALTimétrie, d’Orbitographie et de localization précise/Data Unification and Altimeter Combination System and distributed by the Archiving, Validation and Interpretation of Satellite Oceanographic data, with support from the Center National d’Etudes Spatiales and the Copernicus Marine Environment Monitoring Service. The ADT data has a spatial resolution of 0.25 × 0.25 and is available daily after October 1992. Therefore, we use summertime ADT in 1993–2019 for further analysis.

Besides, the accuracy of altimetry data in coastal areas is still questionable due to land contamination, imprecise tidal corrections, and incorrect removal of atmospheric effects (Risien & Strub, 2016). These uncertainties occur usually in the 20–50 km coastal zone (Vignudelli et al., 2019), so we have masked the nearest grid point to the coast as missing value to avoid these uncertainties.

Due to the coincidence between the geoid and the oceanic surface at rest, measurement of ADT provides a possible way to estimate the surface geostrophic current based on the geostrophic balance between the pressure gradient force and the Coriolis force. At the level surface where density and gravity are essentially constant, a geostrophic current is proportional to the slope of ADT. That is, the surface geostrophic velocities are calculated through the following expression.

\[
(U_x, V_y) = \frac{f}{g} \left( \frac{\partial ADT}{\partial y} \cdot \frac{\partial ADT}{\partial x} \right)
\]
where $U_g$ and $V_g$ are the zonal and meridional components of surface geostrophic velocities, respectively, $g$ is gravitational acceleration, and $f$ is the Coriolis parameter. The Coriolis parameter $f$ is defined as twice the vertical component of the Earth's angular velocity, $\Omega = 7.292 \times 10^{-5}$ rad/sec, multiplied by the sine of the latitude ($y$), that is, $f = 2 \Omega \sin(y)$.

### 3. Quantitative Parameters

As reviewed in the introduction, the upwelling off southeastern Vietnam could appear in either the coastal area or the offshore area. Therefore, we, for the first step, average summertime SST zonally over a 1-degree band along the Vietnamese coastline to check the location of coastal upwelling, and meridionally over the latitudinal band of 10°N–15.5°N from the coast eastward to 116°E to examine the occurrence of offshore upwelling (see the ranges for averaging in Figures 2a and 2d).

After examining the zonally and meridionally averaged summertime SST year to year in 1982–2019, we find that two typical patterns can be identified for both averaged SST (Figure 2). Along the Vietnamese coast, the first pattern has only one trough occurring in the southern part (10°N–12°N) (i.e., one upwelling center; Figure 2b). The second pattern possesses two troughs with one peak in between (i.e., two upwelling centers; Figure 2c) and the southern trough is sharper. The result indicates the along-coast upwelling centers can be grouped into the southern and northern coastal regions separated by 12.5°N, where the local maximum SST is. In the meridionally averaged sense (Figures 2e and 2f), one can also see two typical patterns as found along the coast, that is, one upwelling center versus two upwelling centers. In Figure 2e, only one SST minimum shows up around 108.5°E. Separated by 110°E, two local minimal SSTs appear at ~108.5°E and ~112.5°E with a sharper trough in the west, showing the upwelling in the coastal region is intenser than in the offshore region (Figure 2f). In brief, upwelling centers off the Vietnamese coast can be grouped in three sub-regions (as marked by three dotted areas enclosed by dashed black lines in Figure 1): one is the Southern Coastal Upwelling (SCU; 10°N–12.5°N and 106°E–110°E), another is the Northern Coastal Upwelling (NCU; 12.5°N–15.5°N and 108°E–110°E), and the other is the Offshore Upwelling (OU; 10°N–15.5°N and 110°E–115°E). Based on the property of parameters introduced as follows, each parameter will be applied to all/certain of the three sub-regions.

#### 3.1. SST-Based Upwelling Index

In the literature, thermal difference ($\Delta$SST) between a coastal SST and an offshore reference SST, which is far from the influence of upwelling, has been commonly used as a simple SST-based upwelling index (SSTUI) to quantify the strength of coastal upwelling. The coastal SST is usually defined as the minimal SST recorded in the shelf area, whereas the reference SST is determined in various ways. For example, Nykjær et al. (1994), Santos et al. (2005), Lathurilère et al. (2008), and Marcello et al. (2011), defined the SSTUI as the SST difference between the colder coastal water and the warmer offshore water along a latitudinal section to represent the cross-shore SST gradient. Su et al. (2013) calculated the offshore-onshore temperature difference in the direction perpendicular to the coastline in the northern SCS. Demarcq (1998) used the climatological SST for the reference SST instead of the time-varying offshore SST.

In this study, we establish an adaptive SSTUI for the upwelling phenomenon near the Vietnamese coast based on the following noticeable features. First, the Vietnamese summertime upwelling can spread in a large area covering the SCU, NCU, OU, or two/all of them. Also, after originating in the SCU, the upwelled cold waters could further expand in various directions (such as eastward or northeastward). As a result, signatures of coastal and offshore upwellings frequently co-occur, challenging the determination of the reference SST, in which influences of upwelling and land effect caused by bounding of islands are
considered insignificant. Second, SSTs over the SCS basin suffers strong interannual climate variability (Chao et al., 1996b; Dippner et al., 2007; Fang et al., 2012; Kuo et al., 2004; Li et al., 2017; Park & Choi, 2017; Shu et al., 2016; Wang et al., 2002; Xie et al., 2003; Zhao & Tang, 2007), that is, background SST in the SCS basin may be greatly leveled up/down by climate variability. Thus, using a fixed (or constant) reference SST value, such as the climatological SST, could ignore the influence of the interannual change of intrinsic SST over the SCS basin, resulting in an underestimate/overestimate of SSTUI.

Considering these reasons, we build up an adaptive SSTUI with a time-varying reference SST ($T_{\text{ref}}$) to take into account the influence of intrinsic basin-scale SST. Thus, the SSTUI is calculated by subtracting the minimal SST in an upwelling area ($SST_{\text{min}}(t)$) from the time-varying $T_{\text{ref}}$, i.e.,

$$SSTUI(t) = T_{\text{ref}}(t) - SST_{\text{min}}(t)$$

where $T_{\text{ref}}(t)$ is calculated by averaging summertime SST in the area enclosed by the solid and dashed red lines in Figure 1. In this manner, climate variabilities coexisting in $SST_{\text{min}}$ and $T_{\text{ref}}$ cancel each other out in
the SSTUI. To avoid influences from the land, only SST in the deep water where depth is greater than 500 m (contoured by red dashed lines in Figure 1) is adopted to calculate the $T_{ref}$. In the calculation, the SSTUI has to be positive if upwelling occurs, and the higher SSTUI physically represents the stronger upwelling. As comparing the SSTUI and SST anomaly images in Figure 3, we set 0.25°C as the threshold in the SSTUI time series to represent the appearance of upwelling in the SCU, NCU, and OU; that is, the upwelling occurs as the SSTUI is greater than 0.25°C.

Figure 3. Summertime (June–September) SST anomaly in 1982–2019. Based on the NOAA OISST data, the SST anomaly is derived by subtracting the time-varying $T_{ref}$ (described in Section 3.1; labeled in the upper-right corner) year by year. Red, yellow, and black dots denote the locations of SST minima in the SCU, NCU, and OU, respectively. Black curves denote the central positions of eastward-flowing jet. NCU, Northern Coastal Upwelling; OU, Offshore Upwelling; SCU, Southern Coastal Upwelling; SST, sea surface temperature.
3.2. Wind-Based Upwelling Index

Wind forcing has been regarded as an important factor driving upwelling in the coastal area. Therefore, we apply a wind-based upwelling index (WUI), which is proposed by Gómez-Gesteira et al. (2006), along the Vietnamese coast to examine whether the wind forcing favors the generation of coastal upwelling in the NCU and SCU or not. The WUI is defined as the Ekman transport component in the direction perpendicular to the coastline.

\[ WUI = \sin(\theta)Q_x - \cos(\theta)Q_y \]  

(3)

where \( \theta \) is the angle between the coastline and the due east, and \( Q_x \) and \( Q_y \) are the zonal and meridional components of Ekman transport, respectively. Note that positive (negative) WUI corresponds to an upwelling-favorable (unfavorable) condition (Gómez-Gesteira et al., 2006). The Ekman transport is calculated as follows:

\[ \left( Q_x, Q_y \right) = \frac{1}{\rho_w f} (\tau_y, -\tau_x) \]  

(4)

where \( \rho_w = 1,025 \text{ kg m}^{-3} \) is the density of seawater. Along the Vietnamese coast between 10°N and 15°N, the angle, \( \theta \), varies between 29° and 120°.

3.3. Central Position and Intensity of the Summertime Eastward-Flowing Jet off the Vietnamese Coast

Another important factor regulating the summertime Vietnamese upwelling is the summertime eastward-flowing jet. Because the summertime eastward-flowing jet is predominantly contributed from the geostrophic current (Li et al., 2014), we employ the ADT-derived surface geostrophic velocity to investigate the variation of the jet. In order to track the pathway of the summertime eastward-flowing jet off the Vietnamese coast, the center of the jet is calculated based on the conception of "center of mass" as proposed by Hsin and Qiu (2012) and Hsin et al. (2013). With the zonal geostrophic velocity \( U_x \), the central position \( Y_c \) of the summertime eastward-flowing jet is calculated as follows:

\[ Y_c(x,t) = \frac{\int_{Y_{N}}^{Y_{S}} y U_x(x,y,t) dy}{\int_{Y_{N}}^{Y_{S}} U_x(x,y,t) dy} \]  

(5)

where \( Y_N \) and \( Y_S \) are the northern and southern limits of integration, respectively; other notations are same as those used in the previous equations. In this study, the integration is limited in the range of 8°N–16°N and only the positive \( U_x \) is considered to represent the eastward-flowing jet, that is, \( U_x \) with a negative value is ignored in the calculation.

Along the jet's central position, its intensity (INT) is further calculated by integrating the eastward zonal geostrophic velocity in a 3-degree latitudinal band centered at \( Y_c \):

\[ \text{INT}(x,t) = \int_{Y_{c-1.5}}^{Y_{c+1.5}} U_x(x,y,t) dy \]  

(6)

The central position and intensity of the summertime jet between 108°E and 114°E are calculated. We have also tested variant widths of 2–4° in Equation 6, and robust results are obtained.

4. Results and Discussions

4.1. Interannual Change of the Summertime Vietnamese Upwelling

By subtracting \( T_{\text{ref}} \) (defined in Section 3.1) from the SST at each grid point, the spatial distributions of summertime SST anomaly from 1982 to 2019 are shown in Figure 3. Considering the SST minima in the SCU, NCU, and OU, centers of SST are searched out and marked as red, yellow, and black dots, respectively, in
Along the Vietnamese coast, centers in 1982–1984, 1986–1988, 1991, 1993, 1995–1998, 2004, 2005, 2007, 2010, 2011, 2015, and 2017 appeared concurrently in the SCU and NCU, while those in the remaining years were observed only in the SCU. In the OU, centers were found in most of years between 1982 and 2019, except for 1995, 2010, and 2015. Compared with the coastal centers, distribution of the offshore centers was quite diverse.

By means of the SSTUI, interannual changes of upwelling intensity off eastern Vietnam are further discussed quantitatively for the three sub-regions of SCU, NCU, and OU. In the SCU, the mean of SSTUIs was 1.32°C with a standard deviation (STD) of 0.47°C (Figure 4a). Stronger events with SSTUI greater than 1.79°C (mean + 1STD; red bars in Figure 4a) were observed in 1982, 1984, 2011, 2012, 2014–2016, and 2018, whereas weaker events with SSTUI less than 0.85°C (mean − 1STD; blue bars in Figure 4a) took place in 1987, 1991, 1992, 1995, 1998, 2001, 2002, and 2010. In the NCU, there were 19 upwelling events during the 38 years (Figure 4b), and the mean and STD of SSTUIs were 0.67°C and 0.32°C, respectively. The stronger events, which had the SSTUI greater than 0.99°C (red bars in Figure 4b), took place in 1987, 1991, 1992, 1995, 1998, 2001, 2002, and 2010. In the NCU, there were 19 upwelling events during the 38 years (Figure 4b), and the mean and STD of SSTUIs were 0.67°C and 0.32°C, respectively. The stronger events, which had the SSTUI greater than 0.99°C (red bars in Figure 4b), took place in 1987, 1991, 1992, 1995, 1998, 2001, 2002, and 2010. The weaker upwelling events showed up in 1987, 2005, and 2011 with the SSTUI less than 0.36°C (blue bars in Figure 4b). As to the OU, the mean and STD of SSTUIs were 0.61°C and 0.19°C, respectively. Stronger offshore upwelling (SSTUI > 0.80°C; red bars in Figure 4c) appeared in 1984, 1994, 1997, 2000, and...
2018, while weaker upwelling (SSTUI < 0.42°C; blue bars in Figure 4c) occurred in 1983, 1987, 1992, 2001, 2006, and 2007.

4.2. Trends of Upwelling Intensity
Besides the interannual variation evidenced in the SSTUIs of the three upwelling sub-regions, long-term tendencies seem to be able to be detected in their time series. Thus, the linear trends of the SSTUIs are estimated as well (solid lines in Figure 4). On account of the inconsistent number of upwelling events among the three sub-regions (38 events for the SCU, 19 events for the NCU and 35 events for the OU), slopes of linear regression lines are statistically tested for confidence levels, based on the student’s t-test, to scrutinize the significance of an estimated trend.

For the 38-years time series of SSTUI in the SCU, the slope of the linear regression line is around 0.01°C/decade (p-value = 0.04; black line in Figure 4a), indicative of an apparent intensification of upwelling in the SCU over the entire studied period. On the other hand, the SSTUI in the NCU was stronger in the 1980s and 1990s, and weaker in the 2000s and 2010s, leading to a weakening trend of −0.02°C/decade between the 1980s and 2010s although the estimated trend is not statistically significant (p-value = 0.74; black line in Figure 4b). In the OU, the SSTUI does not show a significant linear trend (0.005°C/decade; p-value = 0.85; black line in Figure 4c), although most of the intense upwelling events took place in the early stage of the studied period.

4.3. Dynamical Mechanism
As mentioned above, the southwesterly wind and the eastward-flowing jet play essential roles in governing the generation and variability of the summertime Vietnamese upwelling (Chen et al., 2012; Dippner et al., 2007; Fang et al., 2000a, 2012; Gan & Qu, 2008; Ho et al., 2000a, 2000b; Hu & Wang, 2016; Li et al., 2014; Metzger, 2003; Kuo et al., 2000; Voss et al., 2006; Wyrtki, 1961; Xie et al., 2003). Due to the uneven distribution of the southwesterly wind and the eastward-flowing jet, these factors could play alternative roles in different upwelling sub-regions. Thus, the dynamical mechanisms leading to the variability of the upwelling intensity are analyzed separately for the three upwelling sub-regions.

4.3.1. Coastal Upwelling: SCU and NCU

4.3.1.1. Upwelling in the SCU
Wind forcing: Located in the coastal region, upwelling in the SCU and NCU are supposed to be affected by Ekman transport, which is induced by wind stresses. For this reason, the WUI, which considers both the effects of orientation of coastline and wind-induced Ekman transport, is utilized to examine the upwelling-favorable conditions of the local wind stress along the Vietnamese coast. Figure 5 presents the WUI averaged in a 1-degree longitude band along the Vietnamese coast and the two components (\(\sin(\theta)Q_x\) and \(-\cos(\theta)Q_y\); hereafter WUI\(_x\) and WUI\(_y\)) of WUI. Generally, most of the WUI along the coastal region were positive (Figure 5a), indicative of the alongshore component of the wind stress along the Vietnamese coast supporting the occurrence of summertime Vietnamese upwelling. The WUI in the SCU (south of 12.5°N; 0.3–3.4 m²s\(^{-1}\)) were on average 3 times greater than those in the NCU (north of 12.5°N; 0.1–1.1 m²s\(^{-1}\)), pointing out that the upwelling-favorable conditions were more prevalent along the southern coast of Vietnam. This fact supports the stronger and more frequent occurrence of upwelling events in the SCU than the NCU. Besides, the correlation coefficient between the SCU-SSTUI and WUI averaged over 10°N–12.5°N is about 0.37 (p-value < 0.1).

Figures 5b and 5c reveal that the two components of the WUI (WUI\(_x\) and WUI\(_y\)) performed differently along the SCU although both of them were positive. The WUI\(_x\) varied from 0.1 to 1.8 m²s\(^{-1}\) while the WUI\(_y\) ranged between 0 and 1.9 m²s\(^{-1}\). The WUI\(_x\) dominated and contributed 51%–98% to the total WUI in the northern part of the SCU, whereas the WUI\(_y\) was predominant and contributed 60%–88% in the southern part. The outcome shows that the upwelling in the southern SCU is mainly generated by the meridional component.
of Ekman transport (induced by the zonal wind stress), while the zonal component of Ekman transport (induced by the meridional wind stress) is much more effective in the northern SCU.

To further check the contribution of wind forcing to upwelling intensity, we calculate the spatial correlation coefficients between SSTUI in the SCU (hereafter SCU-SSTUI) with the zonal wind stress (WSU), meridional wind stress (WSV), and wind stress curl. As shown in Figures 6a and 6b, both of WSU and WSV near the southern coast of Vietnam have positive influences on the SCU-SSTUI with correlation coefficients of 0.27–0.4 (p-value < 0.1). These results confirm again the role of wind stress in generating upwelling in the SCU. However, the low correlation coefficients between the wind stress/WUI and SCU-SSTUI imply that the variation of the upwelling intensity in the SCU may not merely be ascribed to the wind-induced Ekman transport.

Recalling Figures 5b and 5c, the spatially nonuniform distribution of the two components of Ekman transport in the SCU indicates that the Ekman pumping, driven by WSC, could be another factor. With zero WSC contour line extending from ~10°N off the southern Vietnamese coast northeastward to ~16°N off the western Philippine coast, a dipole structure of WSC has been observed off eastern Vietnam in the climatology summertime WSC field (Fang et al., 2002; Ho et al., 2000b; Kuo et al., 2000; Metzger, 2003; Shaw & Chao, 1994; Shaw et al., 1999). The positive WSC occupies the northern flank of the zero WSC contour, including the SCU and OU, whereas the negative WSC expands in the southern flank. Figure 6c shows that the WSC has weak to moderate correlations (R = 0.3–0.5; p-value < 0.1) with the SCU-SSTUI. The
SCU-SSTUI positively correlates with the cyclonic WSC along the southern coastal area of Vietnam but negatively correlates with the anticyclonic WSC in the southern half of SCS. In other words, the enhancement of SCU-SSTUI accompanies the concurrent intensification of both the anticyclonic and cyclonic WSC off the southeast of Vietnam. Composites of wind forcing for the greater and smaller SCU-SSTUI events further evidence that the stronger/weaker WSC dipole shows up as the SCU-SSTUI is greater/smaller, that is, the

Figure 6. Correlation coefficients of SSTUI in SCU (SCU-SSTUI) versus (a) u-component of wind stress (WSU), (b) v-component of wind stress (WSV), (c) wind stress curl (WSC), and (f) intensity (INT; solid black line) and central position ($Y_c$; solid blue line) of the eastward-flowing jet. In (a–c), black box denotes the SCU, and dotted area presents where the $p$-value is less than 0.1 in the student $T$-test. The red dashed line in (f) denotes the separating point ($109.375^\circ$E) of the jet. Note that, in (d and e), different color scales are used for positive and negative WSC. Composites of wind stress (arrow) and wind stress curl (shading) for the events with SCU-SSTUI (d) greater than mean + STD and (e) smaller than mean − STD. Composites of surface geostrophic current (arrow) and absolute dynamic topography (ADT; shading) for the events with SCU-SSTUI (g) greater than mean + STD and (h) smaller than mean − STD. In (g), LC and HC denote the locations of low- and high-center ADT, respectively.
upwelling is stronger/weaker (Figures 6d and 6e). These outcomes show that the WSC dipole plays a vital role in controlling upwelling in the SCU.

The summertime eastward-flowing jet:

To clarify the role of the eastward-flowing jet, the variability of the eastward-flowing jet is quantified by calculating its central position ($Y_c$) and intensity (INT) based on Equation 5 and 6 with ADT-derived zonal surface geostrophic velocity. Figure 7 shows the zonal distributions of the mean $Y_c$ and INT (solid lines) and their variability (1 STD away from the mean; dashed lines). In the west of 109.375°E where the variations of both $Y_c$ and INT are smaller, the $Y_c$ is confined along the coastal area while the INT increases as the jet flows northeastward. After losing the support of the coast (east of 109.375°E), the variations of the jet ($Y_c$ and INT) are dramatically amplified; the jet continues flowing to the northeast until 111°E; afterward, it turns to the east when it weakens slightly until 112°E. By considering the geographical distributions of mean states and variations of $Y_c$/INT, the separating position is estimated to be around 109.375°E and the jet could be divided into the upstream segment (108°E–109.375°E) and the downstream segment (109.375°E–111°E).

Linear regression analysis between the INT and possible regulating factors (Table 1) shows that the interannual variability of the jet's intensity in the upstream segment is closely related to the wind forcing involving the anticyclonic WSC ($R = -0.76$; p-value < 0.1) and the alongshore wind stress ($R = 0.7$; p-value < 0.1). In the downstream, change of the jet's intensity is instead more relevant to the low-center ADT ($R = -0.8$; p-value < 0.1) in the northern flank of the jet.

The comparison of INT and $Y_c$ anomalies shows a good relationship between the jet's intensity/position, especially in the downstream segment (Figure 8). The positive INT anomaly corresponds to the negative $Y_c$ anomaly, showing that intensified (weakened) jet tends to shift southward (northward). Correlation coefficients between INT and $Y_c$ in the same segment are −0.48 and −0.74 (p-value < 0.05), respectively, for the upstream and downstream. Moreover, the INT in the upstream segment and the $Y_c$ in the downstream segment have a greater negative correlation of −0.86 (p-value < 0.05). The above results evidence that the meridional shift of the jet in the downstream segment depends strongly on the variation of upstream jet's strength through the inertial effect.

The variations of pathway and strength of the eastward-flowing jet have been analyzed. To further know the relationship between the jet and upwelling in the SCU, the correlation coefficients between the SCU-SSTUI and INT/$Y_c$ are depicted in Figure 6f. It shows that the INT has positive correlation coefficients with the SCU-SSTUI along its pathway, peaking at around 109°E ($R = 0.5$; p-value < 0.1) in the upstream segment. Whereas, the $Y_c$ is negatively correlated with the SCU-SSTUI with significant correlation coefficients of −0.45 to −0.52 (p-value < 0.1) in the downstream segment. This relationship between the SCU-SSTUI and eastward-flowing jet is further confirmed by the composites of ADT-based surface geostrophic flow and $Y_c$.

### Table 1

| Correlation Coefficient Between the INT and Major Factors |
|----------------------------------------------------------|
|              | Alongshore WS$^a$ | Cyclonic WSC$^a$ | Anticyclonic WSC$^b$ | Low-center ADT$^a$ | High-center ADT$^a$ |
| Upstream INT$^a$ | 0.70              | 0.45             | −0.76               | −0.65              | 0.54               |
| Downstream INT$^a$ | 0.53              | 0.37             | −0.56               | −0.80              | 0.37               |

All p-values are less than 0.1.

$^a$Average alongshore WS in 10–12°N. $^b$Average positive WSC in 108–111°E and 10–12°N. $^c$Average negative WSC in 108–110°E and 6–9°N. $^d$Average ADT in the northern flank of the jet (108–112°E and 12–14°N). $^e$Average ADT in the southern flank of the jet (110–114°E and 8–11°N). $^f$Average INT in the upstream segment (108–109.375°E).

Abbreviations: ADT, Absolute dynamic topography; INT, intensity; WS, wind stress; WSC, wind stress curl.
for the stronger and weaker SCU events (Figures 6g, 6h, and 9a). Stronger jet with pathway shifting to the south appears in the composite of stronger SCU-SSTUI events, and vice versa.

As shown in Figures 6g and 6h, the meridional shift of the eastward-flowing jet is also associated with the intensification of the double-gyre structure which is represented by the ADT dipole localized at two sides of the eastward-flowing jet. The cyclonic gyre corresponds to the low-center ADT (LC labeled in Figure 6g) in the north, while the anti-cyclonic gyre corresponds to the high-center ADT (HC labeled in Figure 6g) in the south. As a part of the warm-core anticyclonic gyre in the southern SCS, the northeastward current along the Vietnamese coast is strengthened (weakened) when both the LC and HC are scaled up (scaled down) in stronger (weaker) upwelling events (Figures 6g and 6h). Moreover, the eastward-expansion (vanishment) of the LC in stronger (weaker) upwelling events could further amplify (abate) its southward countercurrent along the Vietnamese coast. Based on numerical model simulation, Hein (2008) proposed a possible upwelling process in winter that, when a southward boundary current is strong enough to have inertial force greater than the gravity, a bathymetry-induced upwelling along the Vietnamese coast can be generated. Although this dynamical process was applied for the upwelling in winter, it may be valid for the summertime as well. That is, an amplified southward countercurrent on the western flank of the cyclonic gyre (LC) could also promote the upwelling in the SCU during summer due to the topographic effect.

In brief, amplified currents in the stronger SCU-SSTUI events could favor the upwelling in the SCU through the following processes. The amplified ADT dipole intensifies both the northeastward western boundary current along southeastern Vietnam and the southward countercurrent along eastern Vietnam. Furthermore, the intensified northeastward current along the Vietnamese coast sets up a more favorable condition for upwelling, that is, leaner isopycnic or isotherm. At the same time, the intensified southward countercurrent could help the topographic upwelling generation due to the inertial effect.
4.3.1.2. Upwelling in the NCU

Wind forcing: Similar to the SCU, wind stresses also favor the appearance of summertime Vietnamese upwelling in the NCU (north of 12°N) but the favorable WUI values are much smaller (Figure 5). The favorable WUI in the NCU is mainly contributed from the WUI_x while the negative WUI_y occupies 12.75°N–15.5°N, indicating that the WSU plays a role in restricting the development of the coastal upwelling in the NCU.

The spatial distributions of correlation coefficients between SSTUI in the NCU (hereafter NCU-SSTUI) and WSU (Figure 10a) or WSV (Figure 10b) show that the NCU-SSTUI has high negative connections with the remote wind stresses in the southern SCS basin (both WSU and WSV; $R = -0.4$ to $-0.7$; $p$-value < 0.1), but is only positively correlated with the local WSV in the northern NCU ($R = 0.4$ to 0.67; $p$-value < 0.1). This outcome evidences again that the upwelling in the NCU is governed by the local eastward Ekman transport. Furthermore, the negative correlations of both the WSU and WSV indicate the weakening of southwesterly summer monsoon in the southern SCS basin could lead to an intensification of the upwelling in the NCU.

In terms of the WSC (Figure 10c), the NCU-SSTUI is negatively/positively correlated with the negative/positive WSC in the central SCS, similar to those for the SCU-SSTUI but with an opposite sign (Figure 6c). That is, the weakening of WSC dipole could be favorable for developing upwelling in the NCU. Furthermore, the inter-comparison among composites of wind field for the stronger, weaker, and nonupwelling events affirms that the upwelling in the NCU gets stronger as the WSC dipole subsides, and it is weaker or even does not appear as the WSC dipole reinforces (Figures 10d–10f).

The summertime eastward-flowing jet: The linkage between the eastward-flowing jet and upwelling intensity in the NCU is demonstrated in Figure 10g. Correlation coefficient of the INT (black solid line) has negative values throughout the pathway of the jet, with the highest correlation ($R = -0.63$; $p$-value < 0.1) near the separating point. The $Y_c$ displays a negative correlation with the NCU-SSTUI in the upstream segment, but a positive correlation with a peak at $\sim 109.6°E$ ($R = 0.51$; $p$-value < 0.1) in the downstream segment. The relationship shows that the weaker (stronger) eastward-flowing jet at the separating point may cause the jet in the downstream to shift northward (southward) due to the inertial effect and further result in strengthening (weakening) upwelling in the NCU.

To further examine the surface flow patterns in the different states of upwelling in the NCU, the ADT-based surface geostrophic flow (Figures 10h–10j) and $Y_c$ (Figure 9b) are compounded for the stronger, weaker and nonupwelling NCU events. One can see that the dipole ADT weakens and the jet in the offshore area moves to the north as the upwelling strengthens in the NCU, and vice versa. In the nonupwelling condition, the HC (LC) strengthens (reduces) a little compared to the weaker NCU events.

4.3.2. Offshore Upwelling: OU

Wind forcing: The statistically significant correlations between the two components of wind stress (WSU and WSV) and the SSTUI in the OU (hereafter OU-SSTUI) are positive in the northern SCS (Figures 11a and 11b). The WSC dipole off the Vietnamese coast, especially the anticyclonic WSC (negative), still influences the OU-SSTUI in ways resembling the SCU-SSTUI but with reduced correlation coefficients of
0.3–0.48 (p-value < 0.1) and −0.3 to −0.42 (p-value < 0.1), respectively, in (109–112°E, 12–16°N) and (106–116°E, 7–12°N) (Figures 11c). That is, the upwelling in the OU gets stronger (weaker) when the dipole WSC enhances (diminishes), supported again by the wind field composites, in which a more enhanced WSC dipole for a stronger offshore upwelling (Figures 11d–11f).

Figure 10. Correlation coefficients of SSTUI in NCU (NCU-SSTUI) versus (a) u-component of wind stress (WSU), (b) v-component of wind stress (WSV), (c) wind stress curl (WSC), and (g) intensity (INT; solid black line) and central position (Yc; solid blue line) of the eastward-flowing jet. In (a–c), black box denotes the NCU, and dotted area presents where the p-value is less than 0.1 in the student t-test. The red dashed line in (g) denotes the separating point (109.375°E) of the jet. Note that, in (d–f), different color scales are used for positive and negative WSC. Composites of wind stress (arrow) and wind stress curl (shading) for the events with NCU-SSTUI (d) greater than mean + STD, (e) smaller than mean − STD, and (f) smaller than 0.25°C. Composites of surface geostrophic current (arrow) and absolute dynamic topography (ADT; shading) for the events with NCU-SSTUI (h) greater than mean + STD, (i) smaller than mean − STD, and (j) smaller than 0.25°C.

The summertime eastward-flowing jet: As revealed in Figure 11g, the OU-SSTUI has a strong correlation with the eastward-flowing jet in the downstream segment, and is positively correlated with the INT
(R = 0.36 to 0.75; p-value < 0.1) but negatively correlated with the Yc (R = −0.37 to −0.61; p-value < 0.1). The result shows that upwelling in the OU is strongly affected by the eastward-flowing jet in the downstream segment, that is, the greater intensity and southward shift of the jet in the downstream segment accompanies the occurrence of stronger upwelling in the OU. Such a relation is reconfirmed by the composites of surface geostrophic velocity (Figures 11h–11j) and Yc (Figure 9c). The strengthened jet swings to the south in the intense upwelling events, whereas the weakened one shifts to the north in the weaker and nonupwelling events. In the stronger OU event, the dipole ADT is reinforced (Figures 11h). As the dipole ADT is getting weaker, the offshore upwelling is getting weaker (Figures 11i). In the case that the LC significantly retreats and the HC expands to the north, the upwelling does not show up in the OU (nonupwelling event; Figures 11j). These outcomes conclude that the existence of the cyclonic gyre is an essential condition for the occurrence of the upwelling in the OU.

Figure 11. Same as Figure 10, but for the SSTUI in OU (OU-SSTUI).
Based on the previous analyses, Figure 12 summarizes the possible processes of wind stress, wind stress curl, eastward-flowing jet, and oceanic gyre in strengthening the upwelling in different sub-regions. The relationships between these atmospheric and oceanic physical factors and SSTUI in the three upwelling sub-regions are also quantitatively displayed in Figure 13. In the SCU, the local along-shore wind stress, resulting in offshore Ekman transport, can produce coastal upwelling-favorable conditions but is not the main controlling factor in the interannual variability of the coastal upwelling (Figures 12a and 13a). The enhancement of the WSC dipole off the southern Vietnamese coast accompanying the strengthening of the ADT dipole, especially the cyclonic gyre in the south, results in the reinforcement and southward shift of eastward-flowing jet, and is responsible for the upwelling intensifying in the SCU. Associated with the amplified cyclonic gyre, the amplified along-shore southward countercurrent could exaggerate upwelling in the SCU as well.
In the NCU, the offshore Ekman transport resulted from the local WSV is one of the crucial generators for the upwelling, while the changes of both WSC dipole and ADT dipole are the others (Figures 12 and 13b). The upwelling is intensified when the southwesterly monsoonal wind in the southern SCS basin and the WSC dipole off eastern Vietnam subside. In this condition, the ADT dipole is depressed with the weakened eastward-flowing jet shifting to the north, leading to the disappearance of the southward countercurrent. Instead, a northward boundary current flows along the Vietnamese coast in the NCU, which establishes a more favorable condition for coastal upwelling, that is, a shoreward uplift of isopycnics/isotherms is built up.

In the OU, the variability of upwelling pronouncedly connects to the change of the WSC dipole, the intensity and pathway of the summertime jet in the downstream, and the double-gyre structure. The oceanic process plays a dominant role in regulating the interannual variability of the offshore upwelling. The strengthening of WSC dipole and the development of the double-gyre structure promote the enhancement of the eastward-flowing jet in-between, further favoring the formation of offshore upwelling. In the case of the extreme retreat of cold-core cyclonic gyre with the northward expansion of warm-core anticyclonic gyre, the offshore upwelling cannot appear. At the same time, the jet turns in a pure north direction and becomes an extension of the western boundary current. Besides, it is noteworthy that the position of OU center is quite scattered and links greatly to the high variant meridional shift of the eastward-flowing jet.

In addition to the correlation coefficient analyses, we further estimate the vertical velocity based on a simplified omega equation to assess the contribution of Ekman transport, Ekman pumping, and vorticity gradients of the flow field to the upwelling (Johnson et al., 1980; Mazzini & Barth, 2013).

\[
\begin{align*}
\mathbf{w}_{\text{total}} &= \frac{\mathbf{r}_{\text{along}}}{\rho_w \mathbf{R}_d} + \frac{1}{2\rho_w f} \left( \frac{\partial \tau_y}{\partial x} - \frac{\partial \tau_x}{\partial y} \right) + \frac{\mathbf{w}_{\text{purl}}}{2f} + \frac{\mathbf{w}_{\text{e}}}{H \beta \mathbf{v}} \mathbf{V} \mathbf{V} \mathbf{V} \mathbf{V} \mathbf{V} \mathbf{V} \mathbf{V} \mathbf{V} \mathbf{V} \mathbf{V} \mathbf{V} \mathbf{V} \mathbf{V} \mathbf{V} \mathbf{V} \mathbf{V} \mathbf{V} \mathbf{V} \mathbf{V}
\end{align*}
\]

where \( \mathbf{r}_{\text{along}} \) is the wind stress parallel to the coast, \( \mathbf{V} = (u, v) \) is the horizontal velocity, \( \mathbf{R}_d \) is the first mode of the radius of deformation and set to 42 km which is the average in the Vietnamese upwelling area (Dippner,
\( \tau_x \) and \( \tau_y \) are the zonal and meridional wind stress, \( \xi = \frac{\partial f}{\partial x} - \frac{\partial \beta}{\partial y} \) is the relative vorticity, \( \beta = \frac{\partial f}{\partial y} \) is the latitudinal rate of the Coriolis parameter change, and \( H \) is the upper layer thickness.

\( H \) is set as the water depth in the shallow area (water depth < 100 m) and 100 m in the deep area (water depth \( \geq 100 \) m) according to Hein et al. (2013), who estimated the average depth of about 80 ± 20 m off the Vietnamese coast. In the right-hand side of Equation 7, the first term is Ekman transport (\( \varepsilon_{ekw} \)) and only applied for the coastal region, the second term is Ekman pumping (\( \varepsilon_{curlw} \)), the third term is advection of relative vorticity gradients of the flow field (\( \varepsilon_{rew} \)), and the fourth term is planetary vorticity gradients of the flow field (\( \varepsilon_{bew} \)). Based on the summertime AVISO surface geostrophic velocity and ERA5 wind stress in 1993–2019, each component of vertical velocity is able to be estimated. Averaged over the 27 years, the latitude-dependent mean values of terms in Equation 7 are depicted in the coastal region (108–110°E) and the offshore region (110–115°E) (Figure 14). In the coastal region, \( \varepsilon_{ekw} \), \( \varepsilon_{rew} \), and \( \varepsilon_{curlw} \) are mostly positive and on the same order, while \( \varepsilon_{bew} \) is negative. The upward \( \varepsilon_{total} \) is contributed primarily from \( \varepsilon_{ekw} \) and secondarily from \( \varepsilon_{rew} \); however, its meridional change agrees well with \( \varepsilon_{rew} \) (Figure 14a). In the offshore region, the meridional fluctuation of \( \varepsilon_{total} \) is controlled by \( \varepsilon_{rew} \) (Figure 14b). \( \varepsilon_{total} \) possesses significant negative values (downwelling) in the south of 12.5°N while negative and small positive \( \varepsilon_{total} \) appear alternatively in the north of 12.5°N and the upward movement is mainly contributed from \( \varepsilon_{rew} \). Besides, the maximal positive \( \varepsilon_{total} \) in the SCU, NCU, and OU are \( \sim 9 \times 10^{-3} \), \( \sim 4 \times 10^{-3} \), and \( \sim 0.01 \times 10^{-3} \) cm/s, respectively. The relative magnitude among the vertical velocities in the three sub-regions is consistent with the results of SSTUI in which the upwelling in the SCU is much stronger than in the other two areas.

The magnitude and temporal change of estimated vertical velocities agree well with those reported in the literature (e.g., Hein et al., 2013; Wang et al., 2013; Wu et al., 1998). For example, Wang et al. (2013) suggested \( \varepsilon_{ekw} \) of \( 3.6 \times 10^{-3} \) cm/s in the coastal region and maximum \( \varepsilon_{curlw} \) in the offshore region of \( 1.6 \times 10^{-3} \) cm/s. Wu et al. (1998) found a positive vertical velocity of \( 12.5 \times 10^{-3} \) cm/s based on the numerical simulation in 1992–1995. Hein et al. (2013) have conducted numerical simulation and obtained vertical velocities of \( 8 \times 10^{-3} \) and \( 7 \times 10^{-3} \) cm/s in summers of 2003 and 2004, respectively. Although past studies have reported vertical velocities off eastern Vietnam for a specified period, they only considered the wind-induced component (\( \varepsilon_{ekw} \) or \( \varepsilon_{curlw} \)) or estimated \( \varepsilon_{total} \) based on numerical modeling results. This study, for the first time, assesses the contributions of both wind and ocean current to the total vertical velocity.

**Figure 14.** Estimated latitude-dependent vertical velocity averaged over (a) 108–110°E (coastal region) and (b) 110–115°E (offshore region). These values are averaged over the period of 1993–2019. The black, red, green, and blue solid lines denote \( \varepsilon_{re} \), \( \varepsilon_{be} \), \( \varepsilon_{rew} \), and \( \varepsilon_{ekw} \), respectively. The thick dashed line denotes \( \varepsilon_{total} \). In (a), \( \varepsilon_{total} = \varepsilon_{re} + \varepsilon_{be} + \varepsilon_{rew} + \varepsilon_{ekw} \). In (b), \( \varepsilon_{total} = \varepsilon_{re} + \varepsilon_{be} + \varepsilon_{rew} \).
5. Conclusions

Based on the analyses on satellite sea surface temperature images, the upwelling centers nearby the Vietnamese coast are able to be grouped into three sub-regions: the SCU, NCU, and OU. The upwelling in the SCU appeared in all of the study years (1982–2019) with stronger intensity (2 times larger than the other two), whereas the occurrence and intensity of the upwelling in the NCU and OU reduced significantly. Based on the estimated 38-years trends of the SSTUIs in the three upwelling sub-regions, the upwelling intensity tended to increase in the SCU but decrease in the NCU, and no significant trend is found for the OU.

In the coastal area, the along-shore wind stress plays the major role in favoring the coastal upwelling development via the Ekman transport and the wind stress curl plays a secondary role. Accompanying the enhanced dipole of oceanic gyre in the east of Vietnam, the exaggeration of the cyclonic gyre-associated southward along-shore current is responsible for the upwelling intensifying in the SCU but weakening in the NCU. In the OU, the basic condition for the upwelling development is the well-developed gyre dipole, and the change of cyclonic gyre in the north plays a more important role.

In this study, we have suggested the physical processes for the upwelling changes in the three sub-regions of summertime Vietnamese upwelling in the view that the wind forcing and the eastward-flowing jet cause the variability of the Vietnamese upwelling based on the timing and formation of the upwelling-favorable wind and the eastward-flowing jet. However, we cannot exclude the possibility that the upwelling itself can induce feedback on the local current field and wind conditions. For example, Dippner et al. (2013) had claimed that the Vietnamese coastal upwelling can decelerate the boundary current which causes the current separation. Additionally, the quantitative examination of the individual role of cyclonic and anticyclonic WSC in this study suggests that the WSC dipole plays an important role in governing the upwelling in both the coastal and offshore areas but the anticyclonic WSC is more predominant (red and blue bars in Figure 13). This outcome echoes the finding in Castelao (2012) who stated that a strong SST front caused by an enhanced upwelling near Cape Frio, Brazil, can lead to an intensification of negative WSC anomalies in a positive feedback mechanism. That is, through the air-sea interaction, winds and upwelling can influence each other. Therefore, investigations about the contribution of the upwelling feedback into the relationship between the Vietnamese upwelling intensity and dynamical factors are required in the future. We have provided comprehensive descriptions and discussions on the variability and possible physical processes (relevant to the wind field and ocean current) of the summertime Vietnamese upwelling on the interannual timescale. However, these descriptions and discussions are given on the observational base. Relied on numerical modeling experiments in the future, the proposed dynamical processes will be verified and the interannual changes of vertical structure concerning the upwelling will be able to be further examined.

Data Availability Statement

We would like to thank the NOAA ESRL PSD National Climatic Data Center for providing the OISST (https://www.esrl.noaa.gov/psd/data/gridded/data.noaa.oisst.v2.highres.html); the REMSS for the GHRSSST (ftp://remss.com/SST/daily/mw/v05.0/netcdf/); the ECMWF for the wind data, ERA5 (https://www.ecmwf.int/en/forecasts/datasets/reanalysis-datasets/era5/); the AVISO and the CMEMS for the ADT data (http://www.aviso.altimetry.fr/en/home.html); the NOAA National Geophysical Data Center for bathymetry data (https://www.ngdc.noaa.gov/mgg/bathymetry/relief.html).

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