Atmospheric Responses to Kuroshio SST Front in the East China Sea under Different Prevailing Winds in Winter and Spring

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ABSTRACT

Atmospheric responses to the Kuroshio SST front in the East China Sea under different prevailing winds are examined using high-resolution observations and numerical modeling. Satellite data reveal a significant in-phase relationship between SST and surface wind speed, indicative of ocean-to-atmosphere influence. The atmospheric response varies according to the relative surface wind direction with respect to the SST front orientation. Under the alongfront condition, low (high) SLP anomalies are found on the warmer (colder) flank of the front, accompanied by surface wind convergence (divergence). Enhanced precipitation and frequent cumulus convection appear over the warm Kuroshio, suggesting an atmospheric response extending into the free troposphere. Under the cross-front condition, when the air blows from cold to warm (warm to cold) SST, divergence (convergence) is located directly over the SST front, and its magnitude is proportional to the downwind SST gradient. Under such prevailing winds, the SST front has little effect on the SLP and precipitation.

The Weather Research and Forecasting (WRF) Model is used to investigate the mechanism responsible for the atmospheric adjustment. The results show that under the alongfront condition, large temperature and pressure perturbations in the boundary layer are caused by SST gradients, while stability and turbulent mixing are less affected. By contrast, under the cross-front condition, the perturbations of temperature and pressure are small and shifted downstream, while the SST gradient exerts stronger impact on vertical mixing. The modeling results confirm that the pressure adjustment mechanism contributes more to the atmospheric response under alongfront prevailing winds, while the vertical mixing mechanism dominates the atmospheric adjustment under cross-front winds.

1. Introduction

Air–sea coupling associated with ocean fronts and eddies has different features from that associated with large-scale climate modes, which has recently been revealed thanks to advances in microwave satellite technology enabling high-resolution observations in all-weather conditions. On the basin scale, the correlation between sea surface temperature (SST) and near-surface wind speed is often negative, which can be explained by atmosphere-to-ocean influence via surface turbulent heat fluxes (Wallace et al. 1990; Alexander et al. 2002; Xie 2004). However, a different correlation was found near relatively smaller-scale oceanic frontal regions with large SST gradients (Chelton et al. 2004). Stronger (weaker) surface winds were observed to be collocated with warmer (colder) SSTs, suggesting an active ocean forcing on the atmosphere. Extensive literature suggests that this in-phase relationship is ubiquitous and appears near major ocean fronts ranging from tropical trade wind regime to the midlatitude storm track zone (Chelton et al. 2001; Nonaka and Xie 2003; Vecchi et al. 2004; Tokinaga et al. 2005; O’Neill et al. 2003, 2005; Small et al. 2008; Chelton and Xie 2010; Xu et al. 2010). More recent studies demonstrated that the climate effects of sharp SST fronts may extend into the free atmosphere by triggering deep cumulus convection and anchoring rainbands along the warm current. (Minobe et al. 2008, 2010; Tokinaga et al. 2009; Kuwano-Yoshida et al. 2010; Xu et al. 2011; Sasaki et al. 2012).

Two main hypotheses have been put forth to explain the in-phase SST–wind relationship. In the so-called sea level pressure (SLP) adjustment mechanism (Lindzen
findings of S07 can be verified by realistic atmospheric conditions. The question remains as to whether the dependence on cross-frontal wind speed and the latitude responses of the MABL to SST gradients have strong magnitude and SST are caused by variations of larger-scale atmospheric stability. Over warm water, the MABL becomes unstable, and intensified vertical mixing brings down high winds aloft to accelerate surface wind. Over cold water, this downward mixing of momentum is suppressed because of increased stability. These two mechanisms differ in terms of the relative importance each puts on turbulent mixing and pressure gradients induced by SST anomalies. There has been much debate on which mechanism plays a more dominant role in the atmospheric response. Direct atmospheric soundings and large-eddy simulations tend to support the vertical mixing mechanism (Hashizume et al. 2002; de Szoeke and Bretherton 2004; Tokinaga et al. 2006; Skyllingstad et al. 2007), while the results obtained from some high-resolution regional atmospheric models are consistent with the SLP adjustment mechanism (Small et al. 2003, 2005; Song et al. 2006).

Being the western boundary current of the North Pacific, the Kuroshio carries warm/saline water from the tropics and flows northeastward along the continental slope in the East China Sea (ECS). The warm current is separated from the cold shelf water by steep SST gradients: the Kuroshio SST front. The strength of this front is strongest in winter and spring, during which the seasonal mean state of the atmosphere exhibits pronounced impacts exerted by the narrow warm SST tongue west of the Ryukyu Islands and the associated sharp SST front (Xie et al. 2002; Zhang et al. 2010; Xu et al. 2011; Liu et al. 2013).

Recent studies show that air–sea coupling at oceanic mesoscales is closely related to atmospheric background conditions. Based on multiyear measurements from both satellites and moored buoys, O’Neill (2012) and O’Neill et al. (2012) derived an analytical function that shows the observed seasonal and geographical variations in the coupling coefficient between wind stress magnitude and SST are caused by variations of larger-scale ambient wind speed. The idealized experiments by Spall (2007, hereafter S07) suggested that the detailed responses of the MABL to SST gradients have strong dependence on cross-frontal wind speed and the latitude of the front. The question remains as to whether the findings of S07 can be verified by realistic atmospheric conditions or not: in particular, whether or not the atmospheric responses vary with different prevailing winds, and if so, by what mechanisms.

The purpose of this study is to investigate the effects of the Kuroshio SST front on the atmosphere under different wind patterns in the ECS, using a suite of high-resolution satellite measurements, a newly developed atmospheric reanalysis and a state-of-the-art regional atmospheric model. Unlike many other SST frontal zones, the ECS is located in a region where the wind direction exhibits large seasonal variation, making it an ideal place for this study. We put the emphasis on exploring the mechanisms behind the atmospheric responses. The rest of the paper is organized as follows. Data and method are described in section 2. We present an observational analysis in section 3 and explore the simulation results in section 4. A summary and discussion are presented in section 5.

2. Data and method

The SeaWinds microwave scatterometer on board the QuikSCAT satellite measures surface roughness that is converted to the equivalent neutral wind velocity at 10 m. During its lifespan from June 1999 to November 2009, the instrument had revealed rich wind structures on small scales around the world (Chelton et al. 2004; Sampe and Xie 2007). We use the twice-daily product from Remote Sensing Systems (RSS) with a resolution of 0.25° × 0.25° (Wentz and Smith 1999). To make them comparable with other observational data used in this study, we processed them into daily data by averaging the ascending and descending passes of each day. The original daily Advanced Very High Resolution Radiometer (AVHRR) SST data are on a 0.25° × 0.25° grid from the NOAA’s National Climatic Data Center (NCDC) and are available since 1981. Since June 2002, the AVHRR data were merged with the Advanced Microwave Scanning Radiometer (AMSR) microwave measurements to overcome the weakness of infrared instrument that cannot penetrate clouds (Reynolds et al. 2007). The merged AVHRR/AMSR data are used in this study.

We use two sets of the Tropical Rainfall Measuring Mission (TRMM) satellite’s Precipitation Radar (PR) data: 2A25 and 2A23; both are level-2 products distributed by NASA’s Goddard Earth Sciences Data and Information Services Center. TRMM is designed to monitor tropical and subtropical rainfall and has filled many gaps in our understanding of precipitation properties and variation (Houze et al. 2007; Yang and Smith 2008; Romatschke et al. 2010). The 2A25 product provides a 3D orbital precipitation rate with a horizontal resolution of approximately 4.5 km and a vertical resolution of 250 m from surface to 20 km. The 2A23 product provides 2D rain type information by adopting a combination of the horizontal profile method (H method).
and the vertical profile method (V method) introduced by Awaka et al. (1997, 2007). It categorizes precipitation echoes into three types: convective, stratiform, and “the other.” The combination of 2A25 and 2A23 provides us 3D convective and stratiform precipitation rates that are interpolated from an irregular to a $0.1^\circ \times 0.1^\circ$ horizontal grid in this study.

The Modern-Era Retrospective Analysis for Research and Applications (MERRA) is a new atmospheric reanalysis from NASA for the satellite era based on a new version of the Goddard Earth Observing System Model, version 5 (GEOS-5), Data Assimilation System and the NCEP unified gridpoint statistical interpolation analysis scheme (Rienecker et al. 2011). The grid spacing of MERRA is $0.5^\circ$ latitude by $0.667^\circ$ longitude with 72 pressure levels extending to 0.01 hPa. We use this dataset to examine the response of SLP and the vertical atmospheric structure.

All of the above datasets are selected for a common period from 1 December 2002 to 31 May 2009, and are composited to obtain three typical surface wind patterns in the ECS during winter and spring (December–May): namely, the northeasterly wind parallel to the SST front (NE pattern), the northwesterly wind perpendicular to the SST front from cold to warm SST (NW pattern), and the southeasterly wind perpendicular to the SST front from warm to cold SST (SE pattern). The composition analysis is based on daily QuikSCAT vector winds. For example, if more than half grid points over the Kuroshio frontal zone (black rectangle in Fig. 1a) are within the angle range ($\pm 22.5^\circ$) of the NE, NW, and SE pattern winds on a particular day, that day can be counted as a sample for that wind pattern. We obtained 213 samples of the NE pattern, 308 of the NW pattern, and 74 of the SE pattern. Despite the fact that the frequency of SE pattern winds is much less than the other two, a total of 74 samples is adequate to reveal typical air–sea coupling features.

### 3. Observational analysis

#### a. Surface wind speed

First, we briefly describe oceanic conditions in the ECS (Fig. 1). The northeastward Kuroshio leaves a narrow warm tongue in the SST field west of the Ryukyu Islands. Over the shallow continental shelf northwest of the warm current, however, a cold tongue forms under intense surface cooling during the cold seasons. The small heat content of shallow water makes surface cooling more effective. As a result, a sharp SST contrast of more than $10^\circ$C between these warm and cold tongues is evident at $123^\circ$–$127.5^\circ$E. Additionally, the southeastward-directing cold tongue is flanked by two warm tongues, similar to the SST climatology of January–March. Xie et al. (2002) attributed this SST structure to the ocean bottom topography, although their study is based on the SST data in January–March.
From a general view of Fig. 1, the surface wind on the warmer flank of the Kuroshio SST front is 2–2.5 m s\(^{-1}\) stronger than on the colder flank. This in-phase relationship between SST and surface wind is indicative of ocean-to-atmosphere forcing. Upon a closer inspection, the wind speed structures have small but discernible differences among the three wind patterns. For example, the response of surface wind to SST variation from cold to warm and that to SST variation from warm to cold is not symmetric. Good spatial consistency between SST and wind speed exists under the NW pattern winds. Nevertheless, the wind speed minimum of the SE pattern is not perfectly coincident with the cold tongue but slightly shifted to the upwind side because of the excessive deceleration over the SST front.

To isolate the frontal-scale features that are of interest here, we performed high-pass spatial filtering on surface wind and SSTs by removing the 7.5° zonal and 6° meridional running average. High-pass-filtered wind speed anomalies basically overlie SST anomalies of the same sign (Fig. 2), suggesting a tight coupling between SST and surface winds on frontal scale. The positive wind speed anomaly over the Kuroshio between Taiwan and Japan is the largest in the SE pattern and the smallest in the NW pattern, and the negative wind speed anomaly over the cold tongue is small in the NW pattern as a result of the overall stronger wind.

As in many previous studies (Song et al. 2009; Chelton and Xie 2010; O’Neill et al. 2010a, 2012), the mesoscale responses of surface winds to SST are statistically quantified in the binned scatterplots of perturbation wind speed as a function of the perturbation SST for each wind pattern (Fig. 2, right). The slope of the straight line denotes the amount of wind speed change per unit SST variation and serves as an indicator of air–sea coupling strength. Consistent with previous studies, wind speed perturbations on the frontal scale are related linearly to, and correlated positively with, SST perturbations under three wind conditions. The SE pattern wind is most sensitive to SST spatial variation but shows relatively poorer spatial correlation with the SST itself. The opposite is found in the NW pattern, in which the air–sea coupling strength is slightly weaker but the spatial SST–wind correspondence is much better.

Notably, the strong negative wind speed perturbations on the colder side of the front in the SE pattern (Fig. 2e) may, to some extent, because of the difference of equivalent neutral wind speed relative to actual wind speed. Based on in situ buoy observations, O’Neill (2012) found that differences between 10-m neutral wind speed and actual wind speed depend on both stability and wind speed, with larger difference under low wind speed and stable conditions. This effect may thus amplify the coupling coefficient in the SE pattern, but we cannot quantify it because of the lack of actual wind speed observations. Compared with coupling coefficients derived from annual data over other frontal zones (O’Neill et al. 2010a, 2012), the overall coupling strength over the ECS in this analysis is slightly larger than that over the Gulf Stream and the Kuroshio Extension but much smaller than that over the Brazil–Malvinas Confluence region and the Agulhas Return Current region. As suggested by previous studies, the geographical differences in coupling strength are likely as a result of geographic differences in the MABL vertical structure and large-scale forcing.

### b. Divergence, SLP, and precipitation

The Kuroshio SST front also leaves a clear signature in surface wind divergence. In contrast to the overall similar spatial distributions of surface wind speed among the different wind patterns, the divergence depends largely on the orientation of surface wind direction with respect to the SST front. Wind convergence (divergence) is found on the warmer (colder) flank of the Kuroshio SST front as the air flows along the SST isotherms (Fig. 3a). In particular, the wind convergence follows the warm main tongue that meanders from the northeastern tip of Taiwan all the way to the southeast coast of Japan. However, the largest surface wind divergence (convergence) is located right over the strongest SST gradient as the wind moves perpendicular to the SST front from cold to warm (warm to cold) SST (Figs. 3b,c). The high-pass-filtered wind divergence pattern is generally in agreement with the original wind field. In the NE pattern, the perturbation convergences are colocated with positive SST anomalies, and the perturbation divergences are colocated with negative SST anomalies. In the NW pattern, the perturbation field is weaker and shifted downstream, with divergence over the SST front and convergence southeast of the warm tongue. In the SE pattern, strong perturbation convergence is located over the Kuroshio SST front, flanked by weak divergence on both sides.

The disparity of oceanic heating between cold and warm water, on the one hand, tends to form low- (high-) pressure anomalies that cause surface winds to converge (diverge) on the warmer (colder) flank of the SST front. On the other hand, it also changes the intensity of vertical mixing, and thus surface wind speed, causing divergence or convergence directly over the SST front as the air flows across the front. Consequently, the divergence pattern can be viewed as a manifestation of the relative response of pressure perturbation and vertical mixing to the SST gradient. In fact, several studies have revealed that surface wind divergence is linearly
proportional to the downwind component of the SST gradient on the frontal scale under the vertical mixing mechanism (Chelton et al. 2001; O’Neill et al. 2005; Frenger et al. 2013).

Figure 4 presents scatterplots of high-pass-filtered surface wind divergence versus the downwind SST gradient. Perturbation divergence is clearly related to the perturbation downwind SST gradient with a positive correlation of 0.54 in the NW pattern and 0.58 in the SE pattern. The correlation coefficient drops to 0.46 under the NE pattern winds, which may reflect the decrease in relative contribution of vertical mixing in
determining surface wind divergence. The region of near-zero perturbation SST gradients (such as over the cold and warm SST tongues) in the NE pattern shows the least distinct correlation, suggesting other physical processes may dominate surface wind response there. The effects of vertical mixing mechanism can also be seen in the scatterplots between surface wind speed and near-surface stability. In the NW pattern, the wind

FIG. 3. (left) QuikSCAT surface wind divergence (shadings; 10⁻⁵ s⁻¹) and AVHRR SST (contours at the interval of 1°C). (right) As in (left), but for spatially high-pass-filtered fields. (a),(b) NE; (c),(d) NW; and (e),(f) SE wind pattern composites during the winter and spring of 2002–09.
speed is linearly related to the near-surface instability represented by SST minus air temperature at 2 m (SAT) with a slope of 0.71. The slope declines to 0.49 in the SE pattern, but the positive correlation can still be found over most parts of the ECS, except for a stable region of SST – SAT ≤ –1°C. Greater scatter can be found in the NE pattern, and the slope drops further to 0.38.

Minobe et al. (2008) developed a quantitative procedure to measure the pressure adjustment mechanism by examining spatial coherence between SST, SLP Laplacian, and surface wind convergence. This method
has been employed by many researchers to test the SLP mechanism in various oceanic frontal regions using observational data as well as numerical simulation outputs (Bryan et al. 2010; Shimada and Minobe 2011; Nelson and He 2012). Figure 5 presents the spatial patterns of SLP Laplacian calculated from MERRA and their correlation with wind convergence. The Laplacian operator acts as a high-pass filter, extracting the SST frontal effect on SLP that is masked by large-scale circulation. Positive (negative) SLP Laplacian indicates

FIG. 5. (left) MERRA SLP Laplacian (shadings; 10^{-9} Pa m^{-2}) and AVHRR SST (contours at the interval of 1°C); (right) point-to-point comparison between surface wind convergence (10^{-5} s^{-1}) and SLP Laplacian (10^{-9} Pa m^{-2}). (a),(b) NE; (c),(d) NW; and (e),(f) SE wind pattern composites during the winter and spring of 2002–09.
pressure that is lower (higher) than the surrounding mean pressure. In the NE pattern, the low (high) SLP perturbation is distributed along the Kuroshio SST front on the warmer (colder) side, roughly collocated with the surface wind convergence (divergence) zone. In the two cross-front wind patterns, there is no obvious spatial coherence between SLP Laplacian and SST. It is interesting to note that, in the NW pattern, significant low SLP anomalies form over the northeastern part of the warm Kuroshio tongue south of Kyushu Island. Also note in Fig. 1b and Fig. 3c that wind vectors are aligned approximately parallel to SST isotherms in this region, corresponding to locally enhanced surface wind convergence. It is somewhat analogous to the situation in NE pattern; that is, a warm SST tongue produces wind convergence via decreasing surface pressure. Point-by-point comparisons indicate that the wind convergence is linearly related to SLP Laplacian with a slope of 0.32 in the NE pattern. Most data points in the NW pattern are located in the divergence regime, but they correspond to both positive and negative SLP Laplacian, which cannot be expected from the SLP adjustment mechanism. The inconsistency with pressure adjustment also occurred in the SE pattern, in which most data points correspond to negative SLP Laplacian but they spread almost equally between the convergence and divergence regimes.

Previous studies have shown that the SLP mechanism does more than cause MABL adjustment. A narrow rainband with frequent cumulus convection is often observed following the warm SST tongue, accompanied by surface wind convergence and enhanced ascent in the midtroposphere. These results suggest that the SLP mechanism can lead to a deep response extending well into the free atmosphere (Tokinaga et al. 2009; Kuwano-Yoshida et al. 2010; Xu et al. 2011). Figure 6 presents surface total precipitation and the frequency of convective rainfall occurrence. The effect of the Kuroshio SST front on cumulus convection is closely related to the direction of prevailing wind. For instance, a high-frequency occurrence band of convective rainfall closely follows the Kuroshio warm tongue in the NE pattern. Similarly, the total precipitation rate is also markedly modulated by the SST. In particular, a rain rate as large as 10 mm day\(^{-1}\) is observed on the warmer flank of the Kuroshio SST front, in contrast to less than 0.5 mm day\(^{-1}\) over cold water in the northern Yellow Sea. The enhanced convective activity and precipitation over the warm tongue is highly correlated with surface wind convergence, indicative of a deep atmospheric response in the NE pattern. The vertical structure of precipitation rates and vertical velocity (not shown) suggests that precipitation response and upward motion can reach up to a height of 4–4.5 km. Although atmospheric response in this case does not penetrate the whole depth of the troposphere, the influence of the Kuroshio front extends much higher than the MABL top.

Nonetheless, less rainfall occurs over the ECS on the whole in the NW pattern, because of poor moisture content, with large precipitation rate and high convective frequency confined off the east coast of Japan, where the SSTs are relatively low. In the region south of Kyushu, the NW pattern shows more frequent convective precipitation than the NE pattern for the reason discussed earlier. In the SE pattern, the convective frequency is noisy and appears to have no clear correlation with the SST. The total precipitation is mainly concentrated on the colder flank of the SST front where convective precipitation rarely occurs, indicating it is exclusively from stratiform precipitation. The SE wind favors stratiform precipitation over cold water via increasing surface layer stratification and water vapor content by warm and moisture advection.

c. Vertical structure of MABL

The above section shows that the Kuroshio SST front induces large SLP perturbation under alongfront surface winds, while a weak response of SLP is observed as the air flows perpendicularly to the SST front. Actually, over the oceanic frontal regions, the surface pressure perturbation primarily results from pressure variation within the MABL rather than from that in the free atmosphere, as suggested by both flight observations and modeling studies (McGauley et al. 2004; Small et al. 2005). Thus, the vertical structure of the MABL is crucial for the SLP response to SST variation. Figure 7 presents vertical cross sections of high-pass-filtered virtual potential temperature from MERRA along line AB shown in Fig. 3. The 3D high-pass-filtered fields are obtained by applying the same approach to filter surface winds at each vertical level. Significant temperature perturbations are produced by SST gradient in the NE pattern, with positive perturbations giving rise to low SLP anomalies on the warmer flank of the SST front and negative perturbations giving rise to high SLP anomalies on the colder flank of the SST front. In the NW pattern, however, temperature anomalies are relatively small on both sides of the SST front and show a downwind tilt with height over warm water. Similar to the NW pattern, the temperature response in the SE pattern is also shifted to the downwind side, but the thermal effect of the SST front is very shallow, mainly confined to below 950 hPa. Above that height, the anomalies reverse sign, which weakens the
temperature perturbation when integrating over the whole depth of the MABL. This vertical temperature dipole structure reduces the hydrostatic pressure response at sea surface and will be discussed further in section 4.

4. Modeling results
a. Model and experiments

We use the Weather Research and Forecasting (WRF) Model, version 3.5, (Skamarock et al. 2008) to examine...
the atmospheric responses to the Kuroshio SST front over the ECS. To be useful, a model simulation must simultaneously resolve the complex vertical structure of MABL adjustment near the SST front while covering a domain large enough to simulate important physical processes, such as the pressure adjustment. The model domain is 20°–40°N, 114°–140°E. A two-way nest configuration is used with horizontal resolutions of 9 and 3 km for the coarse and fine grids, respectively. A total of 69 sigma levels in the vertical, with 30 levels below 1500 m, is used for both domains. The physics parameterization schemes used in the model and relevant references are listed in Table 1. Since a grid size of 3 km is fine enough to resolve many convective eddies explicitly, we only apply the cumulus parameterization to the coarse domain. The Grenier–Bretherton–McCaa planetary boundary layer scheme is a new physics option in the WRF that has proved effective at producing realistic MABL adjustment to SST variations over the frontal scale (Song et al. 2009).

The initial and lateral boundary conditions are obtained from the NCEP Climate Forecast System Reanalysis (CFSR) dataset, available at 6-h intervals with a horizontal grid spacing of 0.5° and 37 vertical pressure levels from 1000 to 1 hPa (Saha et al. 2010). Over the ocean, the AVHRR SSTs used in the observational analysis are employed here as the lower boundary condition.

We select one case for each of the three wind patterns: the 11 April 2007 NE pattern case, the 12 April 2005 NW pattern case, and the 12 April 2007 SE pattern case. The cases are selected by screening twice-daily QuikSCAT data to ensure that oceanic forcing is the dominant process that modified the observed wind field. To obtain that, we filtered out the observation of which the natural variation in the wind field is large. It is also desirable to select cases with the Kuroshio SST front relatively sharp and straight. We checked the weather map to confirm no atmospheric front was evident near the study area for each selected case. Each simulation involves a 24-h integration during which the SSTs are held constant in time. The model results presented here are those at the end of the 24-h integration, when the MABL nearly

![FIG. 7. Vertical cross sections of spatially high-pass-filtered MERRA virtual potential temperature (shadings; °C) along line AB in Fig. 3a for (a) NE, (b) NW, and (c) SE wind pattern composites during the winter and spring of 2002–09. Spatially high-pass-filtered AVHRR SST (°C) along line AB is also shown at the bottom of each panel.](image-url)
reached a steady state. The starting time of each simulation is chosen so that the ending time is close to that of the QuikSCAT pass. To investigate the impact of the SST front on the overlying MABL, we perform two experiments. The first is the control (CTL) run forced by observed SST, featuring a warm tongue and an associated SST front; the second is the smoothed SST (SmSST) run, in which the SST field is heavily smoothed by applying a two-dimensional 9-point averaging 400 times to remove the warm tongue and the associated Kuroshio SST front.

b. MABL response

In this section, we first test the ability of the WRF Model in simulating the observed in-phase relationship by comparing 10-m neutral winds in the CTL run with those from the QuikSCAT. Then the atmospheric effects of the Kuroshio SST front are presented by the difference field between the CTL and SmSST runs. Figure 8 shows the observed and CTL run surface wind fields. Generally, the surface winds in the CTL runs compare well with the observations. In particular, the model captures the enhanced (reduced) wind speed on the warmer (colder) flank of the Kuroshio SST front. In the NE pattern case, the modeled wind speeds closely correspond to the observations almost over the entire study area. However, a higher wind speed bias of 0.5–1.0 m s\(^{-1}\) over the cold water and a lower wind speed bias of the same magnitude over the warm water in the CTL run can be identified in the NW pattern case, which is possibly due to the weak dependence of turbulent mixing on SST-induced stability variation, as suggested by Song et al. (2009). The model also slightly underestimates the rapid deceleration in the SE pattern case, with a smaller wind speed minimum on the downwind edge of the SST front than the observed one. Despite that, the spatial collocation between SST extreme and wind speed extreme is reasonably well reproduced by the model, especially over the warm tongue.

Figure 9 shows the difference maps of vertically averaged virtual potential temperature, SLP, and SST – SAT. It is noteworthy that the atmospheric responses to the SST anomalies, which are nearly the same in all cases, are quite different under the three wind patterns. In the NE pattern case, both temperature perturbation and its spatial coherence with SSTs are pronounced. Moreover, the same magnitude of SST perturbation on both sides of the Kuroshio SST front yields an asymmetric temperature response, with positive temperature anomalies over the warm tongue larger than the negative anomalies over the cold tongue. The temperature perturbations in the NW and SE pattern cases are generally small and shifted to the downwind side because of the background wind advection. The spatial shift is particularly evident in the SE case, in which the positive perturbation is located on the colder side of the SST front. Consistent with the temperature response, low SLP anomalies are sitting just over the Kuroshio warm tongue, and high SLP anomalies are over the cold SST in the NE pattern case. Additionally, a decrease in pressure at the sea surface over warm water is much larger than an increase in pressure over cold water. The SLP perturbations in the NW and SE pattern cases are less than half of that in the NE pattern case, each with an apparent downwind shift, similar to their temperature perturbations. What is more, the large spatial shift in the SE pattern case makes the high- (low-) pressure perturbations located over the warm (cold) water, a feature opposite to the thermal effect imposed by the SSTs. This type of shift in the SLP field relative to SST due to thermal advection is also discussed by Small et al. (2003) in their numerical studies of the MABL response to Pacific tropical instability waves. The Kuroshio SST front enhances (reduces) near-surface instability on its warmer (colder) flank, yet unlike the responses of temperature and pressure, the SST-induced instability perturbation is most significant in the NW pattern case, followed by the SE pattern case, and is the smallest in the NE case, particularly over the warm tongue. The larger air–sea temperature difference over the warm tongue for the NW and SE pattern cases indicates enhanced surface turbulent heat fluxes and is consistent with a strong vertical mixing mechanism. To explain why the spatial patterns of atmospheric response differ so much under different wind conditions, we further explore the vertical structure of the MABL.

Figure 10 shows vertical cross sections of perturbation virtual potential temperature and turbulence kinetic energy (TKE) along line AB, together with MABL depth for the CTL runs. In the NE pattern case, the MABL is quite shallow over the cold water, with its height close to 100 m, and varies very slowly. It then increases abruptly over the SST front and reaches its maximum of about 650 m over the warm Kuroshio tongue, followed by a gradual decline in the North Pacific. The MABL top is generally higher in the NW pattern case because of the overall greater unstable stratification, but the correspondence between the height and local SST is not perfect, with the maximum height somewhat shifted to the frontal region. There is a drastic change of the MABL top in the SE pattern case, in which the boundary layer collapses over the SST front, from approximately 900 m over the warm tongue to less than 100 m on the cold side of the front, and remains constant thereafter.
FIG. 8. Surface wind speed (shadings; m s\(^{-1}\)) and wind vectors (m s\(^{-1}\)) from (left) QuikSCAT observations and (right) CTL runs in (a),(b) NE; (c),(d) NW; and (e),(f) SE pattern cases. Contours on each panel are for the corresponding SSTs at 1°C intervals. The black rectangle in (b) indicates the fine WRF nest. Note that these plots are for one-day case studies, not the composites of the whole season, as in Fig. 1.
In the model, the virtual potential temperature anomalies in the boundary layer follow the SST anomalies imposed, yet distinct temperature response appears in each case. In the NE pattern case, large positive temperature anomalies are collocated with the warm tongue extending from the sea surface up to 800 m, which leads to an approximately 0.5-hPa decrease in SLP in the same region (Fig. 9b). In addition to the strong cold anomalies below the MABL top via reduced surface heating over cold water, there are positive temperature anomalies riding directly above the cooling effect beneath. This dipole structure of temperature anomalies in the vertical is analogous to those derived from soundings over the eastern tropical Pacific by Hashizume et al. (2002); the authors considered it an alternative interpretation of the “back pressure effect” originally proposed by Lindzen and Nigam (1987). Hashizume et al. attributed the reversed air temperature anomalies near MABL top to the sudden variation of the MABL depth in response to SST anomalies. While both types of back pressure effect emphasized the importance of MABL height variation in reducing SLP perturbation, Hashizume et al. argued that rather than via the additional mass adjustment, as suggested in Lindzen and Nigam’s (1987) model, it was the dipole structure of the temperature anomaly that offset the
SLP effect resulting from SST-induced warming/cooling in the MABL. Obviously, this compensating effect over the cold side of the Kuroshio SST front leads to an asymmetric SLP perturbation in the NE pattern case. In the NW pattern case, the thermal effect of the Kuroshio SST front is well mixed vertically over the full depth of the MABL on both sides of the front. As expected, temperature anomalies are generally small as a result of limited time for MABL adjustment under a strong cross-front wind condition. Additionally, strong advection displaces air temperature anomalies downstream of the SST anomalies, causing downwind shift of SLP perturbation (Small et al. 2003). In the SE pattern case, the cooling effect of cold water is most significant near the surface and is reduced rapidly with height. Above this shallow layer of cold anomalies, the sharp decrease of the MABL top works together with the warm advection by the SE pattern wind to produce large warm anomalies over a deep layer from about 200 to 900 m. Thus, the back pressure effect here is strong enough to reverse the SLP anomalies expected from considering the surface thermal forcing alone. Besides, the positive SLP anomalies over the warm water in the SE case are generated by negative temperature anomalies associated with cold advection.

Although the TKE shows large noisy perturbations at the boundary layer top, turbulent mixing within the MABL is enhanced (reduced) on the warmer (colder) side of the SST front, with the largest anomalies in the NW pattern case and the smallest anomalies in the NE pattern case, a feature that can be seen more clearly over the warm tongue.

Based on the results from 2D idealized simulations, S07 hypothesized that the specific mechanism governing the MABL response was governed by background wind. The vertical structures of MABL response in our experiments are consistent with Spall’s hypothesis and show a clear association with cross-front wind speed. As reviewed by Small et al. (2008), for high cross-front wind speed, such as the NW case presented here, the MABL does not have enough time to adjust to the rapid SST change. On the one hand, large air–sea temperature contrast over warm water markedly destabilizes the MABL and energizes vertical mixing, making the downward momentum transfer more effective; on the other hand, the SLP perturbation tends to be small and shifted downstream. Such a wind pattern favors the vertical mixing mechanism. In contrast, under a weak cross-front wind speed condition, such as in the NE pattern case, the MABL has more time to adjust to the SST gradient; thus, larger air temperature contrast across the SST front enhances the SLP perturbation but simultaneously reduces the perturbation of turbulent mixing because of smaller air–sea temperature difference. With such a wind pattern, the SLP adjustment mechanism tends to play a more important role. In other words, large SST-induced pressure gradient force is expected in the NE pattern case, while a small one is
expected in the NW pattern case because of a strong cross-front wind and deep boundary layer.

c. Momentum budget

We now analyze the momentum budget to further identify the mechanism governing the MABL response. The terms of the momentum equation were output both in the CTL and SmSST runs. The horizontal momentum for the MABL can be written as

\[ \frac{\partial \mathbf{V}}{\partial t} = F_{\text{adv}} + F_{\text{pg}} + F_c + F_{\text{pbl}}, \]

where the terms in the above equation represent, from left to right, the local tendency of momentum, the horizontal advection, the pressure gradient force, the Coriolis force, and the vertical turbulent momentum flux divergence (referred to as the vertical mixing term). Note that we only present the main budget terms, with vertical advection, horizontal mixing, and curvature terms neglected.

To explore how the SST-induced momentum budget influences the surface wind speed, we focus on the difference field of budget terms between CTL and SmSST runs and project them toward the downwind direction, which provides either a driving force (positive) or a drag (negative). Figure 11 shows vertical sections of downwind perturbation pressure gradient (PG) term and downwind perturbation vertical mixing (VM) term along line AB for NW and SE pattern cases. For the NE pattern case, the downwind direction is roughly perpendicular to line AB. To explain how the SST-induced downwind budgets alter winds from cold to warm SST, we present the vertical section for the NE pattern case along the line CD shown in Fig. 8a. In Fig. 11a, the PG term and the VM term are both largest near the surface and decrease with height, but with opposite signs. The PG term reaches maximum over the front and the immediate warm SSTs owing to large baroclinicity over the frontal zone formed by the SST gradients. The negative VM term also shows a local maximum over the SST peak. Thus, the PG term is responsible for the in-phase SST–wind relationship, while the VM term acts in the opposite sense. This is consistent with the “pressure drag” mechanism proposed by Small et al. (2005) and is qualitatively similar to the momentum budget of the cold-to-warm weak wind case in S07. It is apparent that the PG term is much larger than the counterpart in the two cross-front cases. S07 attributed the larger perturbation pressure gradient to the direct result of the weaker background wind allowing the turbulent mixing of temperature in the MABL to act over the same horizontal scale as the front itself. The budget terms in this

Fig. 11. Vertical cross sections of downwind budget terms (Pa m s\(^{-2}\)) for CTL minus SmSST runs along line CD (dashed red) in Fig. 8a for the NE pattern case and along line AB (dashed black) in Fig. 3a for NW and SE pattern cases. Shadings indicate pressure gradient force, and contours denote vertical mixing term for (a) NE, (b) NW, and (c) SE pattern case. The planetary boundary layer height in each CTL run is indicated by a thick dashed black line. The corresponding SST difference (°C) is presented at the bottom of each panel.
analysis confirm the conclusion of S07 that this length scale is determined by cross-front wind speed, for the PG term in the NE pattern case is still dominant for a relatively strong wind but with only a weak cross-front component.

In the NW pattern case, the PG term is relatively small and increases gradually from cold to warm water, with its maximum shifted downstream of the warm tongue. The drag effect of the VM term is diminished over the SST front and the warm SST peak. Considering the PG term shows little change over the same region, the imbalance between the downwind terms accelerates surface flow toward the warmer flank of the SST front. At the top of the MABL, however, the drag by the VM term is enhanced over the SST front. This vertical structure of the VM term corresponds to enhanced vertical momentum flux divergence near the MABL top and reduced momentum flux divergence near surface, suggesting the vertical mixing mechanism actively works to transfer downward momentum from aloft down to the surface by intensified turbulent mixing. Koseki and Watanabe (2010) have split the VM term into two primary parts, indicative of two SST frontal effects included in this budget term. One part represents the vertical mixing mechanism, and the other represents the surface friction against the anomalous flow. If the vertical mixing mechanism works effectively, as in the NW pattern case, the enhanced downward momentum transfer over warm SST offsets the enhanced surface friction, so that the net drag effect of the VM term is reduced for the cold-to-warm flow. Otherwise, the effect of vertical mixing mechanism is overwhelmed by the frictional dissipation so that the net VM term shows an enhanced drag near the surface, forming the pressure-drag balance. Again, it is the cross-front wind speed component that determines how the VM term responds to SST gradients. The momentum budget in the NW case is qualitatively consistent with the 2D idealized simulation of the cold-to-warm strong wind case in S07 and Kilpatrick et al. (2014). It is possible that the vertical mixing mechanism becomes strong enough to even reverse the sign of the VM term near surface, thus driving surface flow to accelerate toward warm water (cf. Fig. 9 of S07). The realistic example of such a VM term acting in concert with the PG term to alter surface wind was found over the Agulhas Return Current region during wintertime (O’Neill et al. 2010b).

In the SE case, the VM term is sharply enhanced at the surface as the air parcel travels from warm to cold SST; whereas at the top of the MABL over the front, the drag by the VM term is reduced. This budget is consistent with the vertical mixing mechanism, corresponding to enhanced (reduced) vertical momentum flux divergence near surface (upper MABL), resulting from less downward momentum transfer due to enhanced stratification. The PG term also changes over the front but not as sharply as the VM term. The PG term in this case cannot be compared with previous studies mentioned above because of the strong back pressure effect, which forms downwind perturbation pressure gradient over the warm water, inconsistent with what is expected from the SST thermal effect.

5. Summary and discussion

During winter and spring, the Kuroshio warm current leaves a visible warm tongue and an associated SST front over the ECS. We focused on the influence of the Kuroshio SST front on the atmosphere in terms of three different wind patterns, using high-resolution satellite observations, MERRA data, and WRF simulation. The results suggest that the frontal effects and the mechanisms by which the atmosphere adjust vary depending on the wind pattern. Stronger pressure adjustment and deeper atmospheric response occurs when the winds blow in an alongfront direction, whereas the vertical mixing becomes a stronger contributor to the atmospheric adjustment under the cross-front wind condition.

The QuikSCAT measured higher wind speed over the warm SST tongue and lower wind speed over the cold SST tongue, indicative of a significant ocean-to-atmosphere impact over the ECS. While this in-phase relationship seems quite robust under different prevailing winds, the divergence field shows a close association with the wind direction. As the wind blows parallel to the SST front, surface wind convergence (divergence) is found along the SST front over warm (cold) SST, accompanied by low (high) SLP anomalies. Further, the TRMM exhibits a deep response extending into the free troposphere, manifested by a narrow rainband with frequent cumulus convection collocating with the warm tongue. When the surface wind blows perpendicularly to the SST front from cold to warm (warm to cold) SST, the divergence (convergence) is located directly over the SST front, with its magnitude positively related to the downwind SST gradient. Meanwhile, there are no clear signatures of the response for either SLP or the convection activity, indicating that the frontal effects are probably confined within the MABL.

The WRF is used to investigate the atmospheric response to the Kuroshio SST front, with a focus on mechanisms responsible for observed effects by the front under three different wind patterns. The CTL runs, forced with high-resolution observed SSTs, reproduce the SST–wind in-phase relationship similar to the satellite observations. Comparison of the CTL and SmSST runs suggests that the thermal forcing of SST anomalies
is most effective under an alongfront wind condition (the NE pattern case), leading to pronounced SLP perturbations collocating with SST anomalies. In contrast, under the cross-front wind condition (NW and SE pattern cases), the thermal effects and resultant SLP perturbations are generally weak and shifted downwind of the SST anomalies by strong advection. Moreover, the “back pressure effect” resulting from the vertical dipole structure of temperature anomalies over the cold water in the SE case gives rise to SLP perturbations with the opposite sign relative to those from direct surface thermal forcing. The Kuroshio SST front also exerts pronounced influence on turbulent mixing in the MABL, with the enhancement of vertical mixing over the warm tongue more significant under the cross-front wind condition. A detailed analysis of the perturbation momentum budget identified a “pressure drag” balance in the NE pattern case, with a large perturbation downwind pressure gradient over the frontal zone contributing significantly to the observed in-phase relationship. However, the same explanation cannot be applied to the NW and SE pattern cases because of the small magnitude of the perturbation pressure gradient along with its phase shift relative to the SST anomalies. Instead, the SST-induced downwind vertical momentum flux divergence term in these cases indicates the vertical mixing mechanism effectively works to alter surface flow. We conclude that the pressure adjustment mechanism exerts a stronger influence on the atmosphere under the alongfront wind condition, while the vertical mixing mechanism dominates the atmospheric response under the cross-front wind condition.

Our WRF results should be treated with caution, because they are derived from a single day simulation that may be a departure from the climatological response. For example, there is the drastic response of MABL temperature and SLP in the SE pattern case simulation (~180° phase shift relative to SST anomaly) and different vertical structures between Fig. 7c and Fig. 10c. Results based on 20 years of MERRA records that contain 250 SE pattern cases show that the climatological phase shift of temperature anomaly is roughly 90°, and the difference between Fig. 7c and Fig. 10c comes from different MABL background wind speed (not shown). The 1000–850-hPa averaged wind speed over the frontal region for Fig. 7c and the WRF SE pattern case is 6.9 and 10 m s⁻¹, respectively. With increasing background winds, the phase shift of MABL temperature and pressure anomalies and the dipolar structure over cold SST in the SE pattern case become more evident.

This study mainly focused on the climatic effects of the Kuroshio SST front. Recently, Liu et al. (2013) found that the pressure gradient effects and near-surface stability effects dominate on different time scales. The SLP adjustment mechanism acts on long time scales, leading to time-mean vector wind strongest over the SST front, whereas the vertical mixing mechanism contributes more at synoptic time scales, causing a time-mean scalar wind maximum over warm SST. According to their analysis, the precondition for the leading role of the thermal wind–SLP mechanism in the governing mean vector wind is that the atmosphere must be quasi-balanced with SST (small cross-front winds) so that large baroclinicity can be formed over the frontal zone. This situation is similar to our NE wind pattern, which confirms the SLP effects on surface wind divergence under alongfront condition. We also followed Liu et al.’s (2013) method to analyze the vertical structure of time-mean vector winds (not shown) and found that the NW and SE wind pattern show a vertical structure more consistent with the vertical mixing mechanism than the NE wind pattern. Moreover, our Fig. 1 suggests that Liu et al.’s (2013) mechanism of scalar wind response works for all wind directions. It is reasonable because their mechanism actually acts on the accumulated kinetic energy, which is irrelevant to wind direction. Therefore, the above mechanism provides an alternative explanation for the in-phase relationship between scalar wind speed and SST in this study. Also, another explanation for the in-phase relationship in the NE wind pattern may come from the simple one-dimensional (1D) balance under a quasi-balance condition proposed by Samelson et al. (2006), in which surface stress magnitude is enhanced over warm SST owing to deeper MABL depth.

It is interesting to note that convection and precipitation in Fig. 6 for the NE wind pattern also occur over a larger area east and south of the Kuroshio, where neither low SLP anomalies, nor surface wind convergence exist. One possible explanation is that the relationship between convection and surface wind convergence via pressure adjustment may only apply to the frontal scale.

The present work focuses on the SST frontal effects on the overlying atmosphere; in reality, the air–sea coupling involves two-way processes. The atmospheric responses to the Kuroshio SST front may feedback to the ocean through the following processes: the in-phase relation between SST and wind speed tends to form a negative thermal feedback by releasing more (less) sensible and latent heat from the ocean over the warmer (colder) SSTs. Another negative feedback may arise from the precipitation and cloud responses. The enhanced cloudiness over the warm current under certain conditions will reduce solar radiation at sea surface and thus damp the SST anomalies. In addition, the small-scale surface wind curl and divergence features
generated by the SST anomalies may significantly affect ocean circulation, as suggested by Chelton et al. (2004, 2007), which will in turn alter the SST anomalies. Since our study reveals that the atmospheric response varies depending on prevailing wind direction, a natural question is whether or not the feedback to the ocean will also differ, and if so, by what mechanism. This requires that full air–sea coupling, rather than just one-way coupling, should be explored next.

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