Two Types of Surface Wind Response to the East China Sea Kuroshio Front*

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ABSTRACT

Effects of the sea surface temperature (SST) front along the East China Sea Kuroshio on sea surface winds at different time scales are investigated. In winter and spring, the climatological vector wind is strongest on the SST front while the scalar wind speed reaches a maximum on the warm flank of the front and is collocated with the maximum difference between sea surface temperature and surface air temperature (SST – SAT). The distinction is due to the change in relative importance of two physical processes of SST–wind interaction at different time scales. The SST front–induced sea surface level pressure (SLP) adjustment (SF–SLP) contributes to a strong vector wind above the front on long time scales, consistent with the collocation of baroclinicity in the marine boundary layer and corroborated by the similarity between the thermal wind and observed wind shear between 1000 and 850 hPa. In contrast, the SST modulation of synoptic winds is more evident on the warm flank of the SST front. Large thermal instability of the near-surface layer strengthens temporal synoptic wind perturbations by intensifying vertical mixing, resulting in a scalar wind maximum. The vertical mixing and SF–SLP mechanisms are both at work but manifest more clearly at the synoptic time scale and in the long-term mean, respectively. The cross-frontal variations are 1.5 m s$^{-1}$ in both the scalar and vector wind speeds, representing the vertical mixing and SF–SLP effects, respectively. The results illustrate the utility of high-frequency sampling by satellite scatterometers.

1. Introduction

The East China Sea is one of the largest shelf seas of the world, surrounded by the Asian continent to the west, the Yellow Sea to the north, and Ryukyu Islands to the east. With the vast continental shelf in the west, the deep Ryukyu Trough runs from southwest to northeast to the west of the Ryukyu Island chain (Fig. 1b). The Kuroshio, a western boundary current of the North Pacific, enters the East China Sea from the east of Taiwan and flows along the continental slope northeastward. The East China Sea Kuroshio (ESK) brings warm/saline water from the tropics, forming a sharp sea surface temperature (SST) front especially in winter and spring when it meets the cold continental shelf water produced by successive cold surges of the East Asian winter monsoon. The ESK front markedly influences local precipitation, clouds, and surface wind (Xie et al. 2002; Zhang et al. 2011; Xu et al. 2011; Sasaki et al. 2012).

Recently, air–sea interaction near SST fronts received much attention (see Chelton et al. 2004; Xie 2004; Small et al. 2008; Chelton and Xie 2010). There are two primary mechanisms for sea surface wind response to SST fronts: vertical mixing effect and SST front–induced sea level pressure (SLP) adjustment (SF–SLP). In the vertical mixing mechanism, warm (cold) SST weakens (strengthens)
the near-surface static stability and thereby enhances (reduces) vertical mixing in the marine atmospheric boundary layer (MABL; Wallace et al. 1989; Hayes et al. 1989), leading to an in-phase relationship between SST and surface wind speed caused by the downward transport of momentum from aloft. Indicative of ocean forcing the atmosphere (Xie 2004), the positive correlation between SST and sea surface wind speed is ubiquitous near major currents ranging from the Kuroshio and its extension (Nonaka and Xie 2003), the Gulf Stream (Chelton et al. 2004), and tropical instability waves (TIWs; Hashizume et al. 2002) to the Brazil–Malvinas confluence in the South Atlantic (Tokinaga et al. 2005). The vertical mixing effect is supported by direct atmospheric soundings (Hashizume et al. 2002; McGauley et al. 2004) and numerical experiments (Song et al. 2009; Koseki and Watanabe 2010).

Lindzen and Nigam (1987) proposed that SLP adjusts to an SST front through the changes in the MABL temperature field. The strong SLP gradient across an SST front results in surface wind change. This SF–SLP effect explains the wind convergence over bands of warm SST (Xie et al. 2001; Minobe et al. 2008), which sometimes anchors deep convection and frequent high clouds (Minobe et al. 2008, 2010; Xu et al. 2011). In contrast to the vertical mixing effect, wind anomalies due to the SF–SLP adjustment are 90° out of phase with SST anomalies in space.
In the ESK region, the prevailing wind is westerly at 850 hPa but northerly at the surface from January to May. The vertical mixing effect cannot explain this change in wind direction. Based upon QuikSCAT data, Zhang et al. (2011) found that the local maximum of surface vector wind is situated over the ESK front and that the SLP adjustment to the SST front is responsible for the northeasterly winds in April [see also Chen et al. (2003)]. In contrast, the maximum scalar wind speed is collocated with the SST warm tongue, suggestive of the vertical mixing mechanism (Xu et al. 2011). Vector wind drives ocean circulation, and the wind convergence may anchor deep atmospheric convection (Minobe et al. 2008; Xu et al. 2011; Sasaki et al. 2012). Scalar wind, on the other hand, is important for surface turbulent fluxes and ocean mixing. Taken together, the above studies suggest different mechanisms for vector and scalar winds but do not explain how and why this happens.

The present study investigates why the vector and scalar winds feature different spatial patterns and what mechanisms are behind the spatial distributions. We show that vector wind is nearly in geostrophic balance with the SLP gradient, which is in turn maintained by the SST gradient of the ESK front. Scalar wind, on the other hand, is related to accumulated wind energy of all frequencies, to which energetic synoptic disturbances are a major contributor. Near-surface atmospheric stability strongly modulates synoptic wind speed across the ESK front, giving rise to the observed scalar wind distribution. By examining the time mean and synoptic perturbations separately, our analysis highlights two different processes (the SF–SLP and vertical mixing effects) that contribute to SST–wind interactions. Previous studies of SST frontal effects based on satellite scatterometry use mainly monthly-mean fields, with a few exceptions such as the work on high-wind occurrence (Sampe and Xie 2007; Cheng et al. 2011). Our results illustrate the utility of high-frequency sampling by satellite scatterometers in studying air–sea interaction at the synoptic time scale.

The rest of the paper is organized as follows. Section 2 introduces the datasets and methods used in this paper. Section 3 describes the major features of the observed scalar and vector wind fields and the effects of the ESK front. Sections 4 and 5 investigate the SF–SLP and vertical mixing effects in the time mean and at synoptic time scales, respectively. Section 6 is a summary with discussion.

2. Data and methods

a. Data

We use high-resolution data to investigate local air–sea interaction around the narrow ESK front (about 200–300-km wide). The monthly climatological–mean vector and scalar winds are calculated based on daily QuikSCAT observations on a 0.25° grid (Liu 2002; http://www.ssmi.com) that resolves sea surface wind variations across SST fronts (Chelton et al. 2004; Xie 2004; Chelton and Xie 2010). QuikSCAT wind velocity is high-pass filtered in time based on 13-day running means in order to examine sea surface wind response at the synoptic time scale. The monthly-mean AVHRR SST on a 0.25° grid from the National Oceanic and Atmospheric Administration (NOAA) National Climatic Data Center (NCDC) is used to track the ESK front (Reynolds and Chelton 2010).

We employ the monthly-mean SST, surface air temperature (SAT), surface wind velocity, and speed from ICOADS at the NOAA Physical Science Division (PSD) on a 1° × 1° grid (Woodruff et al. 2011). Integrating the sea surface meteorological and marine measurements from ships, buoys, coastal stations, and other marine platforms, this dataset has revealed rich structures of atmospheric response to SST fronts and mesoscale eddies (Tokinaga et al. 2005, 2009; Tokinaga and Xie 2009; Tanimoto et al. 2011) and is used to confirm the results derived from satellite observations.

The Climate Forecast System Reanalysis (CFSR; Saha et al. 2010) of the National Centers for Environmental Prediction (NCEP) with a 0.5° × 0.5° horizontal resolution at 6-h intervals is used to investigate air–sea interactions at the synoptic time scale by removing the 13-day running mean. The monthly CFSR climatology is spatially high-pass filtered to reveal the MABL response to the ESK front. In addition, the near-surface thermal stability is calculated from monthly-mean air temperature at 2 m and SST based on the 1-h forecast of CFSR.

b. Methods

The monthly-mean vector wind is defined as follows:

\[
\mathbf{V}_{\text{vector}} = \frac{1}{N} \sum_{i=1}^{N} (u_i \mathbf{i} + v_i \mathbf{j}) = \mathbf{m} + \mathbf{n},
\]

(1)

where \((u_i, v_i)\) is the observed instantaneous wind velocity, \(N\) is the number of observations during a month, \(\mathbf{i}\) and \(\mathbf{j}\) represent the east- and northward unit vectors, respectively, and the \((\mathbf{m}, \mathbf{n})\) denotes the averaged wind velocity over the period. In sections 3 and 4, we analyze the 10-yr climatology of vector wind velocity from January to May 2000–09.

Similarly, the monthly-averaged scalar wind speed (scalar wind hereafter) is calculated according to

\[
V_{\text{scalar}} = \frac{1}{N} \sum_{i=1}^{N} \sqrt{u_i^2 + v_i^2}.
\]

(2)
We use the 10-yr climatology of scalar wind from January to May 2000–09 in sections 3 and 5. We employ a high-pass filter with 13-day windows to extract synoptic perturbations, and the results are robust when the windows are changed from 11 to 30 days. The observed wind \( V_i \) can be decomposed into monthly-mean flow \((\bar{u}, \bar{v})\), synoptic perturbations \((u_0, v_0)\), and perturbations between 13-day and monthly means \((u_\mu, v_\mu)\):

\[
V_i = (\bar{u} + u_0 + u_\mu) + (\bar{v} + v_0 + v_\mu).
\]

The monthly-averaged wind kinetic energy (KE) can be written as

\[
KE = \frac{1}{2N} \sum_{i=1}^{N} (u_i^2 + v_i^2)
\]

\[
= \frac{1}{2N} \sum_{i=1}^{N} [(\bar{u} + u_0 + u_\mu)^2 + (\bar{v} + v_0 + v_\mu)^2]
\]

\[
= \frac{1}{2} V_{vector}^2 + EKE + EKE_e,
\]

where

\[
EKE = \frac{1}{2N} \sum_{i=1}^{N} (u_0^2 + v_0^2)
\]

is eddy kinetic energy (EKE) at the synoptic time scale, \((1/2)V_{vector}^2 = (1/2)(\bar{u}^2 + \bar{v}^2)\), and \(EKE_e\) is the energy between synoptic and monthly time scales. The \(EKE_e\) about 10% of the KE (not shown), is small.

3. Responses of surface vector and scalar winds to the ESK front

The ESK front varies with the season. The SST gradient of the front exceeds \(2 \times 10^{-5} \text{C m}^{-1}\) from January to May and is sharpest in March and April (Fig. 2a). We focus on boreal winter and spring (from January to May) when the ESK front is best developed between the warm Kuroshio and cold continental shelf water in the East China Sea (Bao et al. 2002). On the warm flank of the ESK front, a warm tongue associated with the ESK stretches from the northeast of Taiwan to the south of Yakushima Island (Fig. 1a). A weaker warm tongue originates from the southeast of Cheju Island (33.3°N, 126.5°E) and extends into the central Yellow Sea. To the northwest of the ESK front, there is a cold tongue associated with the shelf water cooled by the East Asian winter monsoon (Xie et al. 2002).

The climatological vector wind from QuikSCAT (Fig. 1a) shows the prevailing northerlies from January to May, as part of the East Asian winter monsoon, with the maximum speed right over the ESK front to the northwest of the warm tongue (i.e., 90° out of phase with the SST maximum). This 90° out-of-phase relationship between the vector wind and SST maxima is consistent with the results of Zhang et al. (2011) for April.
Scalar wind tracks the mean kinetic energy over a period [Eqs. (2) and (4)] and is usually larger than the vector wind speed (note the different color shading scales of Figs. 1a,b) because of the high-frequency variability. Scalar wind speed shows a positive correlation with SST in space with the maximum of wind speed anchored by the warm tongue (Fig. 1b). The in-phase relationship between SST and surface winds is consistent with Xu et al. (2011), indicative of the vertical mixing effect (Xie 2004).

The cross-frontal transections along line AB in Fig. 1d of vector wind velocity, scalar wind speed, and SST – SAT vary in response to the annual cycle of the ESK front (Fig. 2). We employ a monthly standardization (see Fig. 2 caption for details) to reduce the month-to-month variation in order to focus on the spatial variance following the method of Zhang et al. (2011). In the East Asian monsoon domain, the prevailing winds are northerly from September to March and southerly in June and July. The northerlies prevail above the ESK front from November to May when the ESK front is strongest (Fig. 2a). The standardized scalar wind shows an obvious positive correlation with SST – SAT, especially when the ESK is strong (Fig. 2b).

The different relationships between vector and scalar winds and the ESK front (in phase and 90° out of phase, respectively) are compared and further clarified in Fig. 3. Vector wind speed reaches the maximum over the SST front, reflecting the SST frontal effect on SLP (Zhang et al. 2011). The scalar wind maximum, on the other hand, occurs about 1°–1.5° southeast of the ESK front and is collocated with the SST warm tongue, indicative of the vertical mixing effect (Wallace et al. 1989). In addition, SST – SAT reaches a positive maximum at the SST warm tongue. The close collocation of positive SST – SAT and scalar wind speed maxima indicates that the near-surface instability is a direct factor modifying the strength of vertical mixing. Upon close inspection, the wind response is sharper in QuikSCAT than ICOADS especially along the cold flank of the ESK front, most likely because of the coarser resolution and lower sampling frequency of the latter.

Near-surface instability and the ocean current associated with the ESK also affect the wind velocity estimated by QuikSCAT, because it measures the wind stress on the moving sea surface, from which sea surface wind is estimated, assuming neutral instability. The wind velocity difference of QuikSCAT due to the stability dependency of the drag coefficient is one order of magnitude smaller than the observed wind variability across the front (Small et al. 2003; O’Neill et al. 2005). The ocean current speed is about 1 m s\(^{-1}\); much smaller than the 2.5 m s\(^{-1}\) scalar wind speed difference across the ESK front (Fig. 3). QuikSCAT may falsely overestimate the northeasterly vector wind over the northeastward-flowing ESK (about 200–300-km wide), which might weaken the cross-frontal vector wind difference. The in situ observations of ICOADS corroborate the results derived from the satellite data (Figs. 1c,d).

According to Eq. (4), the KE can be approximately decomposed into the kinetic energy of the monthly-mean vector wind and EKE. Therefore, the differences between vector and scalar wind speeds are due to high-frequency weather disturbances that interact with the ESK front. The surface wind responses to the ESK front in the time mean and at synoptic time scale are discussed further in sections 4 and 5, respectively.

### 4. Vector wind response in the time mean

In the ESK region, the prevailing winds are northerly below 950 hPa and westerly above 850 hPa (Fig. 4a). A spatial high-passed filter with a 10° latitude \(\times\) 30° longitude window is employed to extract the local structure in MABL associated with the ESK front from January to May. The air temperature is about 1.5\(^\circ\)C higher over the warm tongue than the cold flank of the SST front at 950 hPa, and the cross-frontal temperature difference increases downward to 8.5\(^\circ\)C at 2 m (Fig. 4b). Thus, a band of pronounced baroclinicity forms over the ESK front in MABL where northeasterly anomalies increase from 1.2 m s\(^{-1}\) at 850 hPa to 2.5 m s\(^{-1}\) at 1000 hPa. The

![Cross-frontal Sections](http://journals.ametsoc.org/doi/abs/10.1175/JCLI-D-12-00092.1)
collocation of the SST front and baroclinicity in MABL suggests weak cross-frontal thermal advection in the time-mean field. The vertically accumulated baroclinicity effect in MABL is the thermal wind, causing the anomalous speed to peak at the surface over the ESK front (Fig. 4b).

Thermal wind in the MABL near the ESK front approximately equals wind shear when ageostrophic flow at 1000 hPa $V_{1000}$ is small (Holton 1979):

$$V_{1000} - V_{850} = V_f + V_{a1000} = V_f$$

$$= \frac{R}{f} \int_{850}^{1000} (V_pT \times k)d\ln P,$$  \hspace{1cm} (6)

where $V_{1000}$ and $V_{850}$ are winds at 1000 and 850 hPa, $V_f$ is the thermal wind between the two levels, $V_pT$ is the horizontal temperature gradient, $f$ is the Coriolis parameter, and $R$ is the gas constant for dry air. The flow at 850 hPa features large-scale variations without a strong SST effect (Fig. 4c), suggesting that the baroclinicity in MABL has a strong effect on surface wind. We integrate the thermal wind downward to evaluate the baroclinicity effect on surface wind, a deviation from the usual applications of the thermal wind concept in tropospheric dynamics. In MABL, the forces of the pressure gradient, Coriolis, and friction are roughly balanced (not shown). The thermal wind resembles actual wind shear where the baroclinicity in MABL is strong. Elsewhere, they disagree as the wind shear is due to the surface friction force acting on the tropospheric pressure gradients.
Based on spatially high-pass-filtered CFSR data and Eq. (6), the actual wind shear and thermal wind between 1000 and 850 hPa are calculated and compared (Fig. 4d). The strong northerly wind shear (black vectors in Fig. 4d) occurs above the ESK front. The thermal wind (white vectors in the right panel of Fig. 4d) tends to be in parallel with the SST contours and strengthens near the ESK front. The similarities in direction and magnitude between the thermal wind and observed wind shear indicate the importance of the baroclinicity in MABL in forming the strong northeasterly winds at the sea surface over the ESK front. The wind shear vectors systematically turn anticlockwise from the thermal wind over the SST front because of the frictional Ekman spiral effect. The cross-frontal component results in the wind shear convergence and divergence on the warm and cold flanks of the SST front, respectively. The significant difference between the thermal wind and wind shear away from the front suggests that other mechanisms such as the surface friction are more important than the baroclinicity.

The geostrophic wind is nearly orthogonal between 850 hPa and the surface, indicating that the surface wind direction is dominated by the baroclinicity in MABL (Fig. 5). While geopotential height at 850 hPa lacks cross-frontal structures, the climatological SLP in winter and spring (Fig. 5a) shows a steep horizontal gradient over the ESK front with strong surface geostrophic wind (Fig. 5b). The Coriolis force cannot be ignored with the large-scale SLP gradient, and the spatial pattern of thermal wind supports the SF–SLP mechanism proposed by Lindzen and Nigam (1987).

Tanimoto et al. (2011) found that the vertical mixing effect accelerates (decelerates) the climatological eastward ageostrophic winds to the north (south) of the SLP minimum induced by the Kuroshio and Kuroshio Extension in winter. Following their method, we found that the ageostrophic wind is strongest over the ESK front rather than over the ESK (not shown), indicative of the vertical mixing effect.

5. Scalar wind response at synoptic time scale

a. Synoptic perturbations of surface wind

Using daily QuikSCAT and 6-hourly CFSR winds from January to May for 2000–09, a temporal high-pass filter based on a 13-day running mean is employed to separate synoptic perturbations ($u'$ and $v'$) from the mean flow. As discussed in section 2, the strength of the synoptic disturbances can be measured by EKE. Similar to scalar wind, the maximum EKE at 10 m (EKE10m) is collocated with the surface warm tongue (Figs. 6a,b) and with the positive peak of SST–SAT, where the near-surface atmosphere is most unstable (Fig. 6b). The EKE10m based on QuikSCAT is much stronger than that from CFSR, indicating an underestimation by the latter of the surface wind response to the ESK front. Without an obvious imprint of the oceanic front, EKE at 850 hPa (EKE850) is very different in spatial distribution from EKE10m (Fig. 6c). Thus, the impact of the ESK front on EKE is mostly confined in MABL. The ratio between EKE10m and EKE850 (EKE10m/EKE850) illustrates SST modulation of surface EKE (Fig. 6d). The ratio EKE10/EKE850 is below 0.6 along the SST cold...
tongue in the East China Sea and reaches 0.8 above the ESK, suggesting that the near-surface stability affects the EKE10m significantly. The ratio EKE10/EKE850 exceeding 1.0 east of Taiwan is probably due to the topography effect.

EKE10m is dominated by large-scale synoptic perturbations but modified by near-surface instability. To investigate the relationship of EKE10m with synoptic perturbations and near-surface stability, Fig. 7 shows EKE10m as a function of SST − SAT and EKE850 based on CFSR. EKE10m is proportional to EKE850 for any given SST − SAT, indicating that the synoptic perturbations at the top of MABL drive the surface variability. The EKE10m response to the EKE850 is significantly larger over the unstable than the stable sea surface (the right panel of Fig. 7), indicating that the near-surface stability controls the sensitivity of the surface wind to 850-hPa perturbations. In addition, EKE10m increases with the interface instability for SST − SAT > 0°C (the bottom panel of Fig. 7), supporting the vertical mixing mechanism. It is interesting that the EKE10m reaches a plateau for a given EKE850 when SST − SAT exceeds 3°C: a saturation condition in agreement with O’Neill et al. (2010). For large SST − SAT, the surface instability no longer influences the EKE10m.
b. Regression analysis

Synoptic wind velocity is generally in the same direction at 850 hPa and the surface. We regress the temporally high-pass-filtered zonal and meridional winds at 10 m onto their counterparts at 850 hPa (Fig. 8). The wind correlation between 10 m and 850 hPa is uniformly high (not shown) because the 10-m wind is first-order driven by pressure perturbations in the free troposphere. The frontal structure of meridional wind regression is much more complex than the zonal wind regression. 

**Fig. 7.** EKE10m is expressed as a function of SST − SAT and EKE850 (shading and dashed contours; m² s⁻²) derived from monthly sampling in the dashed box in Fig. 1c. (bottom) EKE10m averaged for all EKE850. (right) EKE10m averaged under stable-to-neutral (−2° ≤ SST − SAT < 0.5°C; solid line) and unstable (0.5° ≤ SST − SAT < 3°C; dashed line) conditions. The result is robust when altering the averaging of the area of SST − SAT.

**Fig. 8.** Regression coefficient (shading; m s⁻¹) of 10-m wind onto wind at 850 hPa based on monthly temporal high-pass-filtered CFSR from January to May 2000–09, with climatological SST (contours; CI = 1°C) superimposed: (a) zonal and (b) meridional winds.
clearer than the zonal, probably because the meridional wind perturbations are dominant in this season over the East China Sea. Indeed, the variance of surface wind perturbations in the dashed box of Fig. 1c is 7.9 and 14.5 m$^2$s$^{-2}$ for the zonal and meridional components, respectively. For a 1 m s$^{-1}$ meridional anomaly at 850 hPa, the response at 10 m exceeds 0.75 m s$^{-1}$ on the warm flank of the ESK front but falls below 0.55 m s$^{-1}$ on the cold flank (Fig. 8b) because of enhanced vertical mixing over the SST warm tongue.

The probability density function of the regression coefficient (Fig. 9) shows that the ratio of meridional winds at 10 m to 850 hPa is larger for unstable than stable conditions, indicating that the vertical mixing in MABL is intensified dramatically from the stable to unstable near-surface layer. The 10%–90% range of the ratio is 0.55–0.85 for SST – SAT = 4°C but decreases to 0.4–0.6 for SST – SAT = −1°C. In addition, there is an upper limit of 0.9 m s$^{-1}$ on the regression coefficient.

6. Summary and discussion

In boreal winter and spring, a sharp SST front forms between the warm ESK and cold continental shelf water. We have investigated the surface wind response to the ESK front both in the time mean and at the synoptic time scale based on satellite data, ship observations, and high-resolution reanalysis. The results show that the SF–SLP and vertical mixing mechanisms are clear in the time mean and at synoptic time scales, respectively.

Both the responses of vector wind velocity and scalar wind speed at the surface to the ESK front are evident but with different spatial patterns. Vector wind is strongest over the ESK front while scalar wind speed peaks on the warm flank of the SST front and is collocated with the SST – SAT maximum from January to May. Namely, the scalar wind speed and vector wind velocity are 90° out of phase in space with each other. The phase difference suggests that two different physical processes related to SST–wind interaction are dominant at different time scales.

In the time mean, a band of prominent baroclinicity forms in MABL along the ESK front and is collocated with the northeasterly surface wind jet. The observed vertical wind shear and the thermal wind in MABL based on spatially high-passed data are remarkably similar in spatial pattern, suggesting that the vector wind pattern is largely due to the SF–SLP adjustment that originates from the baroclinicity in MABL.

We have used temporally high-pass-filtered data from CFSR and QuikSCAT to study the surface wind response to the ESK front at the synoptic time scale. The spatial patterns of EKE10m and scalar surface wind speed are similar to each other, both with a clear maximum along the ESK, while EKE850 shows a large-scale structure without an evident imprint of the ESK front. This, together with the maximum in SST – SAT, indicates that the ESK enhances the wind variability at the surface via the vertical mixing mechanism.

The SST front acts as a steady forcing on MABL because of the slow nature of SST variations. The geostrophic adjustment is fast, and vertical mixing adjustment is even faster. They are both at work in the time mean and
the synoptic time scale. The SF–SLP adjustment dominates in the time mean because synoptic vector wind perturbations are averaged out. At the synoptic time scale, on the other hand, the vertical mixing mechanism modulates surface wind more strongly than the SF–SLP mechanism because of the large fluctuations of wind and atmospheric stability. Indeed, Fig. 10 shows that SST–SAT variations are larger at the synoptic time scale than at the monthly time scale. Strong cross-frontal wind and thermal advection cause extreme values of SST–SAT, a process more prominent in synoptic disturbances than in the monthly mean.

The vector wind peaks over the SST front, while the scalar wind maximum is displaced toward the warm flank by 150 km and collocated with the peak of the near-surface instability (SST – SAT). For simplicity, we have attributed the latter stability effect to the vertical mixing of momentum, an assumption that may not always be valid. Small et al. (2003, 2005) show that cross-frontal thermal advection displaces air temperature anomalies in such a way that the resultant SLP–wind anomalies are in phase with SST anomalies, making it difficult to distinguish between the SF–SLP and vertical mixing mechanisms from the phase relationship. The positive curl and convergence of vector wind velocity are collocated with the warm SST tongue (Fig. 11), consistent with the SF–SLP adjustment (Minobe et al. 2008). The surface wind convergence may anchor strong precipitation accompanied by deep convection, enhanced high cloud occurrence, and an increased lightning flash rate over the Gulf Stream in summer (Minobe et al. 2010) and over the ESK in spring (Xu et al. 2011) and early summer (Sasaki et al. 2012).

More in-depth diagnosis into the temperature and momentum budgets is necessary to fully understand the dynamics of wind adjustment to the ESK front. A promising method for such a diagnosis has recently been developed for the Gulf Stream by N. Schneider et al. (2011, personal communication) based on the statistical analysis of high-resolution atmospheric reanalysis. We plan to apply this method to the ESK front in the future to shed light on the dynamical processes.

Previous studies of ocean front–atmosphere interaction tend to focus exclusively on monthly and longer averages. This study showed that the high-frequency sampling of synoptic disturbances by QuikSCAT is a key to understanding the spatial pattern of scalar wind in relation to the ESK front. More such studies that take advantages of satellite temporal sampling are desirable to shed light on scale interactions.

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