Surface Wind Responses to Mesoscale Sea Surface Temperature over Western Boundary Current Regions Assessed by Spectral Transfer Functions

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ABSTRACT: Satellite observations have revealed that mesoscale sea surface temperature (SST) perturbations can exert distinct influence on sea surface wind by modifying the overlying atmospheric boundary layer. Recently, spectral transfer functions have been shown to be useful to elucidate the wind response features. Spectral transfer functions can represent spatially lagged responses, their horizontal scale dependence, and background wind speed dependence. By adopting the transfer function analysis, the present study explores seasonality and regional differences in the wind response over the major western boundary current regions. Transfer functions estimated from satellite observations are found to be largely consistent among seasons and regions, suggesting that the underlying dominant dynamics are ubiquitous. Nevertheless, the wind response exhibits statistically significant seasonal and regional differences depending on background wind speed. When background wind is stronger (weaker) than 8.5 m s\(^{-1}\), the wind response is stronger (weaker) in winter than in summer. The Agulhas Retrification region exhibits stronger wind response typically by 30% than the Gulf Stream and Kuroshio Extension regions. Although observed wind distributions are reasonably reconstructed from the transfer functions and observed SST, surface wind convergence zones along the Gulf Stream and Kuroshio Extension are underrepresented. The state-of-the-art atmospheric reanalysis and regional model represent well the structure of the transfer functions in the wavenumber space. The amplitude is, however, underestimated by typically 30%. The transfer function analysis can be adapted to many other atmospheric responses besides sea surface wind, and thus provide new insights into the climatic role of the mesoscale air–sea coupling.

KEYWORDS: Boundary currents; Marine boundary layer; Air-sea interaction; Mesoscale processes; Sea surface temperature

1. Introduction

High-resolution satellite observations and numerical models have revealed that sea surface temperature (SST) perturbations on a scale of approximately 50–500 km exert distinct influence on sea surface stress and wind. A mesoscale warm (cold) SST perturbation accelerates (decelerates) surface wind, yielding wind divergence and curl (e.g., O’Neill et al. 2003; Chelton et al. 2004; O’Neill et al. 2005; Small et al. 2008; Chelton and Xie 2010). The wind response can be spatially shifted relative to mesoscale SST perturbations because of background horizontal advection of temperature and moisture (Small et al. 2003; Masunaga et al. 2015; Schneider and Qiu 2015; Foussard et al. 2019a). Distinct time-mean surface wind convergence zones have been identified along major western boundary currents, including the Gulf Stream and Kuroshio Extension (e.g., Minobe et al. 2008; O’Neill et al. 2017). Ocean eddies also leave their signatures on the overlying atmosphere (Frenger et al. 2013; Putrasahan et al. 2013; J. Ma et al. 2015; Masunaga et al. 2016).

The regime of the surface wind response changes depending on background wind velocity as characterized by the cross-front Rossby number $\epsilon = U/(fL)$ (Kilpatrick et al. 2014), where $U$ signifies background wind speed perpendicular to an SST front, $f$ the Coriolis parameter, and $L$ the width scale of an SST front. When background wind crosses an SST front ($\epsilon \ll 1$), vertical turbulent mixing and horizontal advection are the dominant terms in the momentum balance and thus “vertical mixing mechanism” plays a dominant role (Wallace et al. 1989; Hayes et al. 1989). When background wind blows along an SST front ($\epsilon \gg 1$) and hence horizontal advection is negligible, air temperature within the atmospheric boundary layer adjusts to SST underneath. The resultant sea level pressure gradients modulate surface wind (“hydrostatic pressure adjustment mechanism”) (Lindzen and Nigam 1987). The regime separation has been explored by numerical experiments with a simple 2D configuration (Song et al. 2006; Spall 2007; Kilpatrick et al. 2014, 2016) and an idealized boundary layer model (Schneider and Qiu 2015).
The influence of mesoscale SST perturbations has been suggested to reach the tropopause level (Minobe et al. 2010; Kurowano-Yoshida et al. 2010; Hand et al. 2014). The signatures of mesoscale SST perturbations can be identified in cloudiness and precipitation as well as surface winds (e.g., Tokinaga et al. 2009; Frenger et al. 2013; Masunaga et al. 2015). Furthermore, the mesoscale air–sea coupling can modulate synoptic-scale systems (Parfitt and Czaja 2016; Vannière et al. 2017; Parfitt and Seo 2018; Foussard et al. 2019b; Masunaga et al. 2020a,b) and basin-scale circulation (Feliks et al. 2004; Révelard et al. 2016). Ocean eddies in the vicinity of the Kuroshio Extension can exert remote influence, affecting precipitation along the western coast of the North America through modulating storm-track activity (X. Ma et al. 2015). Also, potential feedback forcing from the atmospheric responses to the upper ocean have been discussed (White and Annis 2003; Seo et al. 2007b; Small et al. 2008; Perlin et al. 2020).

The strength of air–sea coupling has been quantified by using a coupling coefficient, which is a local regression coefficient of surface wind onto SST perturbations. The coupling coefficient varies regionally and seasonally (Chelton et al. 2004; O’Neill et al. 2005) and is found to be sensitive to model resolution and parameterization schemes (Maloney and Chelton 2006; Song et al. 2006; Seo et al. 2007a). Recently, Schneider (2020) has proposed using spectral transfer functions for quantifying the wind responses (Schneider and Oju 2015). As in a coupling coefficient, spectral transfer functions are linear regression coefficients, but now in the wavenumber space. Spectral transfer functions can represent wind responses that are spatially shifted relative to SST perturbations, and are formulated to be dependent on background wind speed. Furthermore, underlying physical processes can be distinguished since horizontal scale dependence of the wind responses is illustrated as the wavenumber dependence.

Schneider (2020) argues that, in the Agulhas Retroflection region, the wind response is a function of the horizontal scale and the Rossby number, and is consistent with a lee inertial gravity wave interacting with the vertical mixing and hydrostatic pressure adjustment processes. Here, we explore whether these dynamics are active in other western boundary current regions and seasonally modulated by applying the transfer function analysis. We also assess the performance of the state-of-the-art atmospheric reanalysis and a regional atmospheric model in representing the mesoscale air–sea coupling.

The rest of the paper is organized as follows. The data and methodology are described in section 2. In section 3, we investigate seasonality and regional characteristics in transfer functions based on satellite observations. We discuss the reconstruction skills in section 4. In section 5, the performance of an atmospheric reanalysis and regional numerical model is investigated. A discussion and summary are given in section 6.

2. Data and method

a. Satellite observations, an atmospheric reanalysis, and a regional model

We mainly use the ascending track observations of gridded daily zonal and meridional sea surface wind components of Quick Scatterometer (QuikSCAT) version 4 (Ricciardulli et al. 2011) and 3-day running mean of Advanced Microwave Scanning Radiometer for Earth Observing System (AMSR-E) SST version 7 (Wentz et al. 2014). Both are provided by Remote Sensing Systems (RSS) on a 0.25° × 0.25° latitude–longitude grid. For estimating transfer functions, rain-flagged observations are excluded. Data missing regions including the landmasses are filled with interpolation. The analysis period is 2003–08, when QuikSCAT and AMSR-E overlap. We exclude the descending track data to reduce computation time. Nevertheless, we have confirmed that the differences in the annual midlatitude transfer functions discussed in section 3 between the ascending and descending track data are negligible.

For comparison, we also use daily SST and surface wind products of Japanese Ocean Flux Dataset with Use of Remote Sensing Observation version 3 (J-OFURO3) (Tomita et al. 2018). The daily SST product is an ensemble median of at least nine satellite SST products for 2003–08 in each day. The surface wind product is an ensemble average of eight or more satellite products. The products are available on the same grid as the RSS products. Although the AMSR-E and QuikSCAT products are incorporated into J-OFURO3, the comparison between these datasets provides insights into data sensitivity of the transfer function estimation.

To assess the performance of an atmospheric reanalysis in representing the mesoscale air–sea interaction, we use ERA5 atmospheric dataset (Copernicus Climate Change Service 2017; Hersbach et al. 2018). The horizontal resolution is T639 (approximately 31-km resolution) with 137 sigma–pressure hybrid vertical levels. ERA5 assimilates scatterometer surface winds including QuikSCAT. For its lower boundary condition, the Met Office Hadley Centre sea ice and sea surface temperature dataset version 2 (HadISST2) with 0.25° × 0.25° resolution (Hirahara et al. 2016) is prescribed until August 2007, and the Operational Sea Surface Temperature and Sea Ice Analysis (OSTIA) with 0.05° × 0.05° resolution (Donlon et al. 2012) from September 2007 onward. Masunaga et al. (2015, 2018) argue that insufficient resolution in prescribed SST distinctly affect the representation of the atmospheric boundary layer in atmospheric reanalysis. Nevertheless, our visual inspection confirms that SST distribution in ERA5 adequately represents mesoscale features around the western boundary currents, and no obvious inconsistency is identified at the time of the SST data replacement.

We also conducted a set of regional atmospheric experiments with the Weather Research and Forecasting (WRF) modeling system version 4.2.1 (Skamarock et al. 2019). The WRF system has been widely used for mesoscale air–sea coupling investigations (e.g., Song et al. 2009; O’Neill et al. 2010; Kilpatrick et al. 2014). Three-month integration have been carried out over the Agulhas Retroflection, Gulf Stream, and Kuroshio Extension domains for December–February and July–August starting from 25 November and 25 May of each year, respectively. The horizontal resolution is 20-km with 27 vertical levels. For the initial and lateral boundary conditions, NCEP FNL operational analysis (National Centers for Environmental Prediction 2000) is utilized. The lateral boundary is updated every 6 h. OISST is prescribed for the lower boundary
condition and updated daily. We use the Mellor–Yamada–Janjić scheme (Janjić 1994) for planetary boundary layer physics, Thompson et al. scheme (Thompson et al. 2008) for microphysics, and Tiedtke scheme (Tiedtke 1989; Zhang et al. 2011) for cumulus parameterization. For straightforward comparison with QuikSCAT data, one snapshot (0000 UTC) at each day is utilized from the ERA5 and WRF outputs for SST, 10-m zonal and meridional wind components, sea level pressure, and 2-m surface air temperature after interpolating onto a 0.25° × 0.25° grid. For ERA5, 10-m equivalent neutral wind components are also utilized for comparison.

b. Transfer function analysis

We adopt the spectral transfer function approach proposed by Schneider (2020) with some updates to facilitate seasonal and regional comparison. The formulation is briefly described in this subsection and discussed in more detail in the appendix.

Winds are split into large-scale and ocean mesoscale components, U and u, respectively. The ocean-mesoscale winds are represented as wavenumber integral of SST Fourier coefficients (T) multiplied by transfer functions (Å) plus residuals (δu):

\[
u = (2\pi)^{-2} \int d\mathbf{k} \hat{\mathbf{Å}} \hat{T} e^{-i\mathbf{k} \cdot \mathbf{x}} + \delta u,
\]

where tildes signify complex numbers, x location vectors, and k horizontal wavenumbers. On the natural coordinate system, Å consists of downwind and crosswind components, which are aligned with and perpendicular to background winds, respectively. As shown in Fig. 1, Å is dependent on lee wavenumbers (abscissa) that are aligned with the downwind component, and port wavenumbers (ordinate) aligned with the crosswind component; Å is also dependent on background wind speed.

As seen in Eq. (1), the transfer functions correspond to regression coefficients of surface winds onto SST; thus, the unit is m s⁻¹ K⁻¹. The real part illustrates wind responses that are in phase with SST perturbations, and the imaginary part represents 90° out-of-phase responses. When background wind blows along (perpendicular to) an SST front in the physical domain, the situation corresponds to the lee (port) wavenumber being 0 in the wavenumber space.

Transfer functions are estimated from the observations or model outputs by the least squares fitting procedure. Target domains are the Agulhas Retro, Gulf Stream, and Kuroshio Extension regions as enclosed by thick lines in Fig. 2, where surface wind exhibits prominent wind divergence as a response to mesoscale SST perturbations (e.g., Small et al. 2008). For every target grid point, Fourier coefficients of SST are estimated for a rectangular box with 8° longitude × 8° latitude size centered at the target point, and then observed ocean-mesoscale wind components at the target point are fitted using Eq. (1). Daily background wind speed and direction are derived by applying the two-dimensional moving average filter with a radius of 350 km. The corresponding ocean-mesoscale winds are estimated by subtracting the low-pass filtered winds. The target domains shown in Fig. 2 are designed so that more than 80% of the area of every 8° box is covered with sea.

The least squares fitting is carried out on a daily basis throughout the analysis period for all of the target points.

Target seasons are December–February and June–August. We also estimate “midlatitude” transfer functions by combining the three target domains that serves as a reference for comparison [see the appendix and Eq. (A11) for more details]. Crosswind components and crosswind axes are scaled by a factor of fi/fj to take into account the sign differences of the Coriolis parameter between the hemispheres. Thus, we test a hypothesis that rotation plays an important role. If that is the case, the transfer functions should have similar structure between the hemispheres by applying the fi/fj scaling.

To determine the base background wind speeds (U0) for the base transfer functions, wind roses over the target regions are examined (Fig. 3). Over the Gulf Stream and Kuroshio Extension regions, westerlies with typically 14 m s⁻¹ are dominant in winter, whereas weak southerlies with 6 m s⁻¹ are dominant in summer. On the other hand, the seasonality is indistinct over the Agulhas Retrospection region. We chose nine background wind speeds from 1.0 m s⁻¹ with 2.5 m s⁻¹ intervals for estimating base transfer functions.

To estimate statistical significance, we randomly subsampled 100 horizontal observation points from each target domain and then obtain transfer functions for the same analysis period. We repeated the same procedure 25 times; thus, we obtain 25 transfer functions. Although the individual transfer functions exhibit somewhat noisy structure, averages across the 25 transfer functions are nearly identical to those estimated with all the observation point. Therefore, statistical significance is estimated by Student’s t test at the 99% confidence level with these 25 transfer functions by assuming they are mutually independent. The null hypothesis is that amplitudes are zero.

3. Seasonality and regional differences based on the satellite observations

Figure 1 shows annual midlatitude transfer functions for downwind (left two columns) and crosswind (right two columns) responses. The transfer functions are estimated over the three target domains combined, throughout winter and summer. The structure resembles those obtained over the Agulhas Retrospection region detailed by Schneider (2020). The amplitude of the transfer functions is the largest around U0 = 11 m s⁻¹, and gets smaller for either stronger or weaker background wind. For weak background wind (≤3.5 m s⁻¹), the distinct signals tend to be confined to lee wavenumbers smaller than the scale of unit Rossby number (gray dashed lines), where horizontal scale is large or background wind is blowing along an SST front. We argue that the pressure adjustment process mostly yields these responses, since air temperature would adjust to SST under the weak horizontal temperature advection, leading to strong sea level pressure responses through the hydrostatic balance (e.g., Kilpatrick et al. 2014, 2016). On the other hand, under stronger background wind (≥11 m s⁻¹), the transfer functions exhibit larger amplitude at lee wavenumbers larger than the unit Rossby number. With the strong background wind speed, the temperature
FIG. 1. (left two columns) Downwind and (right two columns) crosswind transfer functions (m s$^{-1}$ K$^{-1}$) estimated over the three target domains combined with the winter and summer months for 2003–08 based on QuikSCAT and AMSR-E obtained from RSS. Their real and imaginary parts are shown separately as indicated above each column. The abscissa indicates lee wavenumbers (cycles per 100 km) and the ordinate indicates port wavenumbers scaled with $f/|f|$. The transfer functions are estimated for nine background wind speeds (m s$^{-1}$) as indicated in the individual panels. The vertical dashed lines signify the lee wavenumbers that correspond to the unit Rossby number estimated from background winds. Small dots are applied where the signals are insignificant at 99% confidence level.
The difference between surface air and sea surface can be large due to horizontal temperature advection, leading to modulation in surface winds through the vertical mixing process. These two processes are clearly separated at unit Rossby number for the imaginary part of the downwind response. The real parts of downwind response exhibit prominent peaks at the lee wavenumbers corresponding to unit Rossby number, where vertical mixing and pressure adjustments may interplay.

The transfer functions exhibit asymmetry across the axes. For instance, the fourth quadrant in the real part of downwind responses exhibits smaller amplitude than the first quadrant. In the physical space, these features imply that, under westerly background in the Northern Hemisphere, a southwest–northeast-oriented SST front, like the Gulf Stream over 60°–70°W, yields weaker wind responses than a northwest–southeast SST front. The opposite is the case for the real part of crosswind responses regarding their absolute amplitude. The amplitude of the transfer functions attenuates toward higher wavenumbers, which implies that correlation between mesoscale winds and SST perturbations becomes weaker for smaller horizontal scales. This result is inconsistent with the idealized boundary layer model developed by Schneider and Qiu (2015). Meanwhile, transfer functions based on the WRF outputs exhibit the same features as discussed in section 4, hence these features are not artifacts of the satellite observations.

To investigate seasonality, midlatitude transfer functions are estimated only from winter months or from summer months. As shown with contours in Fig. 4, summertime midlatitude transfer functions are qualitatively similar to the annual ones. Wintertime transfer functions (not shown) exhibit the same features. The similarity suggests that the underlying dynamics are the same regardless of seasons. Thus, the annual transfer functions serve the typical structure and seasonality can be perceived as deviations around it.

The differences between the summertime and annual transfer functions are typically less than 10%, with larger relative changes of 30%. Still, statistically significant seasonal differences can be identified (colored in Fig. 4). When background wind is weak ($U_0 < 8.5$ m s$^{-1}$), the amplitude in summertime transfer functions are larger than that in wintertime counterparts by typically 30%, significant at the 99% confidence level. On the other hand, when background wind is stronger, the amplitude is found to be larger in winter than in summer. The background wind dependence of the seasonality has not been highlighted in previous studies. Individual transfer functions estimated over each target region separately exhibits largely the same seasonality (not shown).

As in seasonal differences, regional differences are found to be relatively small. As shown with contours in Fig. 5 for wintertime Agulhas Retroflection region as one example, the structure of individual transfer functions estimated over each region resembles the annual midlatitude transfer functions. The differences between individual transfer functions and midlatitude counterparts are at most ~30% in either winter or summer. Notice that the similarity in the transfer functions between the hemispheres imply that rotation plays an important role and the difference in the Coriolis parameter signs are appropriately taken into account by using the $f/|f|$ scaling. Nevertheless, as shown with color in Fig. 5, we can identify that the Agulhas Retroflection region exhibits larger amplitude in transfer functions than the other two target domains both in winter and summer (not shown) at the 99% confidence level. The amplitudes...
FIG. 3. Wind roses of daily background wind based on QuikSCAT estimated for December–February and July–August, separately, for 2003–08 over the domains indicated above individual panels. The radial coordinate indicates probability (%); the azimuthal direction shows wind direction. Wind speed (m s\textsuperscript{-1}) is discretized as shown at the bottom.
FIG. 4. As in Fig. 1, but estimated only with summer months (contoured with 0.1 m s\(^{-1}\) K\(^{-1}\) intervals; zero contours are omitted) and their differences from transfer functions estimated only with winter months (colors; summer minus winter). The thick (thin) contours correspond to positive (negative) values. Every other base background wind speed are picked up for visibility. Dots are applied where the differences are insignificant at 99% confidence level.
FIG. 5. As in Fig. 4, but annual transfer functions estimated over the Agulhas Retroreflection domain estimated with December–February and July–August combined (contours) and their difference from annual transfer functions estimated over the Gulf Stream domain (shaded).
are comparable between the Gulf Stream and Kuroshio Extension regions (not shown). This may be because of larger signal-to-noise ratio over the Agulhas Retroreflection region because of less active atmospheric disturbances (discussed in section 4).

4. Reconstruction of the mesoscale wind field

We investigate skills of transfer function based reconstruction to reproduce observed wind distribution. Daily wind responses are reconstructed with the inverse Fourier transform [i.e., first term of the right-hand side in Eq. (1)] by using

![Graphs and charts showing reconstructed wind fields and skill scores for different regions and seasons.]
transfer functions, satellite-observed SST, and background winds. The transfer functions estimated for each region and each season over the 6 years are utilized. The corresponding transfer functions for wind divergence are estimated by Eq. (A6). The reconstructed wind components are averaged over the 6-yr period for each season, and then compared with the corresponding observational averages.

The similarity between reconstructed and observation is assessed by spatial correlations and skill scores. A skill for downwind components \( s_1 \), as an example, is defined as

\[
s_1 = 1 - \frac{\sum x \left[ \mu_1^c(x) - \mu_1^o(x) \right]^2}{\sum x \mu_1^c(x)^2},
\]

where \( \mu_1 \) signifies downwind components of surface winds. The position vector \( x \) covers the individual target domains.

Overbars signify time average. The superscript \( O \) signifies the satellite observation and \( R \) reconstructions. Square of the correlation coefficients corresponds to a lower bound of wind variance explained as a linear response to mesoscale SST, as compared with atmospherically induced variance. Skills for crosswind and divergence are assessed in the same way (Fig. 6). In general, the seasonality in the correlation and skill are indistinct. The downwind response exhibit higher correlation and skill than the crosswind response, probably because more complicated process are dominant in the crosswind response including sea level pressure response and Doppler-shifted near-inertial lee wave (Schneider 2020).

In the Agulhas Retroreflection domain, the reconstructions capture the observed distributions well both in winter (Fig. 7) and summer (not shown), as consistent with the high skills and correlations (Fig. 6). Although atmospheric disturbances may leave distinct imprints on the time-mean wind divergence distribution (O’Neill et al. 2017), the surface wind divergence

![Fig. 7](image-url)
response to SST explains as much as ~50% of the mesoscale spatial variance of the observations. However, the reconstruction underestimate wind divergence features finer than 100 km as consistent with the high-wavenumber attenuation in the transfer functions. It would be worth mentioning that the skills are improved typically by 0.1 by smoothing out the fine-scale structure in the observation with the low-pass filter.

Transfer function based reconstructions overall reproduce the observed wind distribution also over the Gulf Stream region both in winter (Fig. 8) and in summer (not shown). The reconstructions, however, underestimate the zonal band of prominent wind convergence along the Gulf Stream seen in the observation. The underrepresentation is associated with the underestimation in the crosswind dipole around (39°N, 65°–50°W) (Figs. 8c,d). This result is consistent with recent studies that argue the important roles of atmospheric disturbances on time-mean wind divergence patterns (O’Neill et al. 2017; Parfitt and Seo 2018; Masunaga et al. 2020a,b). Further investigation on daily wind divergence confirms that the Gulf Stream region experiences extreme wind divergence.

FIG. 8. As in Fig. 7, but over the Gulf Stream domain for December–February.
events more frequently than the Agulhas Retroreflection region (black and red lines in Fig. 9, respectively). Nevertheless, the reconstructions well represent the observations in the northeastern part of the domain where mesoscale SST exhibits larger amplitude. The linear response explain as much as ~45% of the northeastern part of the observation.

The corresponding reconstruction in summer exhibits the similar characteristics. The divergence reconstruction skill is slightly smaller than in winter, while the atmospheric disturbance activity is weaker (black dash line in Fig. 9). We speculate that this is because spatial variance in high-pass-filtered SST in summer is smaller than in winter (Figs. 6a,e).

Reconstruction skills for the Kuroshio Extension region are poorer, in particular, over the southern part (Fig. 10). The skill in wind divergence reconstruction improves only slightly to be 0.26 for the northern part. The lower skills are probably owing to the weaker mesoscale SST variation (Figs. 6a,e) and frequent extreme divergence events (green lines in Fig. 9). Notice that the Kuroshio Extension exhibits marked decadal-scale variability (Qiu and Chen 2005; Qiu et al. 2014), and SST gradients were relatively weak in its unstable regime for 2006–08 (Wang and Liu 2015; Masunaga et al. 2016). The reconstruction skills may be improved in the stable regime, for which longer data period is needed to obtain sufficient realizations.

We further examine reconstruction skills for shorter time scales. The daily reconstructed wind components are averaged for each month or each season (i.e., 3-month average) of a year, and then compared with the corresponding averaged observations. To obtain robust wind reconstructions, the seasonal transfer functions for each region estimated with the whole 6-yr period shown above are used. The reconstruction skills reduce rapidly because of the reduction of the signal-to-noise ratio for shorter time averaging. For the Gulf Stream and Kuroshio domains in both seasons, the seasonal-mean skills reduce to less than 0.3 for crosswind components and 0.2 for divergence. They further halve on their monthly-mean reconstruction. Meanwhile, over the Agulhas Retroreflection domain in both seasons, the seasonal-mean reconstruction skills retain relatively high values (around 0.35 for crosswind and 0.25 for divergence). Even monthly mean reconstruction skills (correlations) are larger than 0.2 (0.45) for both crosswind and divergence. Reconstruction skills for downwind components reduces in the same way, but typical monthly-mean skills retain typically 0.6, 0.4, and 0.3 for the Agulhas Retroreflection, Gulf Stream, and Kuroshio domains, respectively.

5. Comparison between RSS, J-OFURO3, ERA5, and WRF outputs

In the physical space climatology, ERA5 and WRF appear to well represent the observed surface wind divergence (Fig. 11). In this section, the performance of ERA5 and WRF are assessed by using the spectral transfer functions.

a. Comparison in wind responses

Transfer functions for surface winds based on ERA5 (contoured in Fig. 12) and WRF (not shown) resemble those based on the RSS satellite products, including the scale separation of physical processes by unit Rossby number, background wind speed dependence, and high-wavenumber attenuation, suggesting that underlying physical processes are reproduced. The amplitude is, however, underestimated typically by 30%. The underestimation may be related to under-representation in vertical diffusion (Song et al. 2006).

We also examine the transfer functions based on J-OFURO3 (contoured in Fig. 13). The differences from the RSS satellite products are relatively small as the J-OFURO3 winds are composed of satellite observations. Nevertheless, we found that, in the crosswind component when background wind speed is 6–11 m s−1, the RSS transfer functions (Fig. 1) more extend into the higher port wavenumbers. In the imaginary part of the downwind component, the separation of the signals at unit Rossby number is less clear in J-OFURO3. The differences in the real parts of the downwind response can reach 40%. Thus, the transfer functions capture these nonnegligible differences.

To understand the wind responses more intuitively, transfer functions based reconstructions using Eq. (1) are estimated to an idealized SST pattern. The idealized SST pattern has a Gaussian shape,

\[ T(x) = T_0 \exp\left(\frac{-x^2}{2L^2}\right). \]

where \( T_0 = 1 \) K and \( L = 70 \) km.

Figure 14a shows the reconstructed wind responses to the Gaussian shape SST estimated with the annual midlatitude transfer functions from the RSS satellite products. The background wind is blowing from left to right at 11 m s−1. The
wind response indicates distinct speed increase near the SST peak up to 0.4 m s$^{-1}$. The wind response maximum is slightly shifted into the fourth quadrant. The response pattern also illustrates a mesoscale circulation blowing into a low pressure center near 200 km downwind of the SST peak. The asymmetry across the background wind is probably owing to constructive (destructive) interference of the pressure gradient force and Doppler-shifted near-inertial lee wave in the first (fourth) quadrant (Schneider 2020).

The corresponding reconstructions based on J-OFURO3, ERA5, and WRF (Fig. 14) illustrate largely consistent patterns. However, J-OFURO3 overestimates the wind speed increase near the SST peak, and the downwind extensions of the response are slightly wider in the crosswind direction. In the ERA5 and WRF reconstructions, the wind speed responses near the SST peak is underestimated by 30%. Also, the wind response in ERA5 underestimates negative downwind response around 300 km downwind side and is more elongated into the fourth quadrant. The corresponding wind convergence response is also weaker by 30% (not shown).

Given that the satellite products are calibrated with equivalent neutral wind, transfer functions for equivalent-neutral wind is more relevant for validation. Although the climatological-mean wind divergence are nearly identical between the actual wind and equivalent-neutral wind (Fig. 11), the downwind equivalent neutral wind transfer functions exhibit larger amplitude (not shown), hence the corresponding reconstruction is closer to the satellite counterpart (Fig. 14d). This is a reasonable result because equivalent-neutral wind speed is larger than actual wind speed when the surface layer is unstable (e.g., Liu and Tang...
Samelson et al. (2020) also argue that the coupling coefficient for equivalent-neutral wind is larger than those for actual wind. Therefore, the weaker amplitude in ERA5 and WRF with actual winds may be partly attributable to the difference between actual wind and equivalent-neutral wind. Still, the downwind equivalent-neutral wind response is underestimated by 25% compared with the satellite observations. On the other hand, transfer functions for crosswind equivalent-neutral wind are nearly identical to the actual wind counterpart. These results imply that the crosswind response occurs when air temperature is close to SST and hence the stability is neutral (Liu and Tang 1996), where the pressure adjustment process works efficiently.

b. Responses in sea level pressure and surface air temperature

To obtain more insight into atmospheric responses, the transfer functions for sea level pressure and surface air temperature at 2-m height are examined (Fig. 15). Their real parts exhibit the largest amplitude where lee wavenumbers are close to zero. In the physical space, these correspond to background wind flowing along an SST front, where horizontal advection of temperature is negligible. Under the situation, the boundary layer temperature reflects SST structure underneath, yielding strong pressure responses. When background wind has the component that crosses an SST front, surface air temperature responses shift downwindward relative to SST perturbations due to horizontal advection. Thus, the sea level pressure and 2-m temperature responses are manifested in the imaginary parts for the lee wavenumbers larger than zero. When background wind is 1 m s$^{-1}$, the real part of the 2-m temperature transfer function resembles a Gaussian function. The result implies that the corresponding surface air temperature response in the physical space is similar to spatially low-pass filtered SST by a Gaussian filter (Bracewell 1986; Hamming 1988) (see online supplementary material). The distinct sea level pressure responses are limited to smaller wavenumbers than the 2-m temperature responses, indicating that upper atmospheric temperature needs more time to adjust to SST underneath than 2-m temperature. In other words, cross-front Rossby number [$\epsilon = U/(fL)$] needs to be even smaller as discussed in the introduction. Also, smaller-scale structure would be more susceptible to horizontal dissipation.
FIG. 12. As in Fig. 4, but estimated based on ERA5 for December–February and July–August combined (contours) and their difference from satellite observations (colors). Statistical significance indicators are omitted.
FIG. 13. As in Fig. 12, but estimated based on J-OFURO3.
Transfer functions for the Laplacian of sea level pressure and temperature differences between SST and 2-m air temperature have been also estimated (Fig. 16). See the appendix for the derivation. The temperature difference exhibit distinct positive responses in the real part, indicating that the surface atmosphere is statically unstable on warm SST perturbations. The transfer functions for the Laplacian of sea level pressure largely re\textsuperscript{fl}ect the sea level pressure responses, but emphasize the responses on the higher wavenumbers. The wavy pattern is probably caused by the spectral leakage.

The reconstruction of sea level pressure responses to the Gaussian SST (Fig. 17) exhibits a low pressure center on the downwind side of the SST peak, which extends into downwindward due to background advection as consistent with Foussard et al. (2019a). The pressure structure is consistent with the mesoscale circulation pattern in Fig. 14. Surface air temperature response exhibits warming near the SST peak and a downwindward extension. Surface stability response, which is measured as the SST specified minus air temperature response, exhibits strong destabilization near the SST peak that can act to increase the wind speed. The downwindward extensions become shorter for weaker background wind (not shown). Thus, the reconstructions largely yields physically reasonable patterns. These results confirm that ERA5 and WRF can represent the mesoscale air–sea interaction processes adequately, and the transfer function analysis can be adopted for investigating sea level pressure and surface air temperature response as well as surface wind responses. Note that the sea level pressure response in ERA5 exhibits the crosswindward broadening near 300 km on the downwind side, while that in WRF simply gets weaker downwindward. We have not explored the physical validity of these features. Nevertheless, we have confirmed the broadening is less distinct for 1979–84 in ERA5, when less observations are available for the data assimilation. Thus, we speculate that the feature relates to data assimilation.

6. Discussion and summary

We adopt the spectral transfer function analysis proposed by Schneider (2020) to investigate surface wind responses to...
mesoscale SST perturbations. Seasonal and regional differences in transfer functions are found to be relatively small; thus, annual midlatitude transfer functions represent the typical wind response features. To further confirm these results, we compare the transfer functions by reconstructing the wind responses in the physical space. For a particular season and region, two reconstructions are estimated. One is reconstructed with the transfer functions for the target season and region as done in section 4, and the other uses the annual midlatitude transfer functions. Although the differences in the amplitudes between the reconstructions can reach 20%, their spatial correlations are at least 0.95. These results imply that the underlying dynamics are ubiquitous over the western boundary current regions and seasons. Also, the sign difference in the Coriolis parameter is successfully incorporated by the $\|f\|$ scaling.

Nevertheless, we can identify statistically significant seasonality and regional differences. For weak background wind ($U_0 \approx 8.5 \text{ m s}^{-1}$), transfer functions tend to exhibit larger amplitude in summer than in winter, and vice versa for

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**Fig. 15.** As in Fig. 1, but estimated for (left two columns) sea level pressure and (right two columns) surface air temperature based on ERA5 over the three target domains for December–February and June–August combined. Every other base background wind speed are picked up for visibility. Statistical significance indicators are omitted.
stronger background wind. The larger amplitude in summertime transfer function for weak background wind may be related to deep atmospheric responses mediated by convection (Minobe et al. 2010; Kuwano-Yoshida et al. 2010). On the other hand, distinct atmospheric responses to mesoscale SST is typically limited within the atmospheric boundary layer (Minobe et al. 2010). Thus, the larger amplitude in wintertime transfer functions for strong background wind may be related to the deeper atmospheric boundary layer in winter (Fig. 18), which would lead to larger sea level pressure response and larger momentum transport from higher altitude (Feliks et al. 2004; Foussard et al. 2019a). Besides, the larger sample numbers for the strong background wind in winter (Fig. 3) can lead to larger signal-to-noise ratio. Between the western boundary current regions, the Agulhas Retroreflection region yields the largest amplitude in transfer functions regardless of seasons.

We will explore these transfer function features in our future work with using the linear boundary model developed by Schneider and Qiu (2015). As a preliminary investigation with the linear model, we have confirmed that the $f$ dependence of the wind responses within midlatitudes is relatively weak. Therefore, the differences in the typical latitudes between the target domains cannot yields the statistically significant regional differences.

The observed mesoscale wind distributions are well reconstructed by using transfer functions and observed SST. For
the downwind response, the reconstruction captures as much as 60%–80% of the observed spatial variance. Over the Agulhas Retroreflection region, the wind divergence reconstruction captures as much as 40% of the observation. The wind convergence zones along the Gulf Stream and Kuroshio Extension are, however, poorly represented as consistent with the distinct imprints of synoptic-scale disturbances on time-mean wind divergence (O’Neill et al. 2017; Parfitt and Seo 2018; Masunaga et al. 2020a,b).

In this paper the focus has been on the midlatitude oceanic frontal zones, but the method can be applied in other regions such as the eastern equatorial Pacific. Preliminary transfer function analysis over tropical instability waves revealed that the divergence response was better captured in the reconstruction, possibly because of weaker influence of synoptic atmospheric disturbances. These results are to be presented in future work.

The transfer functions can be utilized to assess performance of numerical models. ERA5 and WRF underestimate amplitude of transfer functions and hence the wind responses, while the horizontal structures are consistent with those based on the satellite observations. ERA5 underestimates the wind responses despite assimilating QuikSCAT. Other reanalysis products may suffer the same issue. The underestimation may be partly attributed to the difference in actual wind and equivalent neutral wind. Still, these results suggest that parameterizations of boundary layer processes need to be improved (e.g., Song et al. 2006). Although the J-OFURO3 and RSS products are both composed of satellite observations, their transfer functions exhibit the structural differences in the wavenumber space. We also demonstrated that the transfer function analysis can be applied to many atmospheric variables besides surface winds.

Transfer and impulse response functions can be utilized to filter out the surface wind features associated with mesoscale SST perturbations. We will conduct sensitivity experiments using the filter to further explore the importance of the mesoscale air–sea coupling by using numerical models. Recently, horizontal resolution of numerical models has been rapidly enhancing, and importance of ocean submesoscale air–sea coupling on the climate system has been suggested (e.g., Lamberts et al. 2013; Ma et al. 2016; Su et al. 2018). Our investigation would extend to ocean submesoscale when observations and models with submesoscale-resolving resolution become more widely available.

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![Figure 17](image-url)
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APPENDIX

Estimating the Transfer Function

This appendix describes how spectral transfer functions are estimated. The transfer functions are estimated on the natural coordinate system using the least squares fitting to minimize the difference between the observed and estimated winds. We refer the readers to Schneider (2020) for a more detailed description.

a. Natural coordinate system

Sea surface horizontal winds \( \mathbf{u}_t \) are decomposed into background and ocean-mesoscale components \( \mathbf{u}_w \) with the

FIG. 18. Climatologies of atmospheric boundary layer depth (m) based on ERA5 for 2003–08 over the Gulf Stream domain in (a) winter and (b) summer months. (c),(d) As in (a) and (b), respectively, but for the Agulhas Retroreflection domain.

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spatial scale of larger and smaller, respectively, than \( O(1000) \) km, namely,

\[
u_i = U \hat{e}_U + u_i, \tag{A1}
\]

where \( U \) and \( \hat{e}_U \) signify speed and unit vectors of background winds.

The ocean-mesoscale wind components are represented in the natural coordinate system, which is spanned by the unit vertical vector \((\hat{e}_3, \hat{e}_U, \hat{e}_i)\), and \( \hat{e}_3 \times \hat{e}_U \); 

\[
u_1 = \hat{e}_U \cdot u, \tag{A2}
\]

and 

\[
u_2 = (\hat{e}_3 \times \hat{e}_U) \cdot u, \tag{A3}
\]

which represent changes in speeds and directions of \( u_i \), respectively. For simplicity, \( u_1 \) and \( u_2 \) are referred to as “downwind” and “crosswind” components, respectively.

In the present study, we evaluate background winds as low-pass filtered surface winds by applying the two-dimensional moving average with a radius of 350 km. The ocean mesoscale wind components and SST with horizontal scale smaller than 750 km are estimated by subtracting their low-pass filtered surface winds by applying the two-dimensional moving average with a radius of 350 km. The ocean mesoscale wind components are represented in the wavenumber space. Mesoscale SST is measured in the wavenumber space. Mesoscale SST with horizontal scale smaller than 750 km are estimated by subtracting their low-pass filtered surface winds by applying the two-dimensional moving average with a radius of 350 km. The ocean mesoscale wind components and SST with horizontal scale smaller than 750 km are estimated by subtracting their low-pass filtered surface winds by applying the two-dimensional moving average with a radius of 350 km.

\[W(t) = \frac{1}{(2\pi)^2} \int d\mathbf{k} \hat{T}(\mathbf{k}) e^{-i\mathbf{k} \cdot \mathbf{x}}, \tag{A4}\]

where tildes signify complex numbers, \( \mathbf{x} \) the location vectors in the local Cartesian system, and \( \mathbf{k} \) the horizontal wavenumbers. Ocean-mesoscale winds \([u = (u_1, u_2)]\) are represented with an integral of SST Fourier components multiplied by transfer functions \([\hat{A} = (\hat{A}_1, \hat{A}_2)]\) over the wavenumbers plus residuals \([\delta u = (\delta u_1, \delta u_2)]\), namely,

\[u = (2\pi)^{-2} \int d\mathbf{k} \hat{A}[\hat{e}_U \cdot \mathbf{k}, (\hat{e}_3 \times \hat{e}_U) \cdot \mathbf{k}, U] e^{-i\mathbf{k} \cdot \mathbf{x}} + \delta u. \tag{A5}\]

Transfer functions are dependent on the lee wavenumber \((\hat{e}_U \cdot \mathbf{k})\), port wavenumber \([ (\hat{e}_3 \times \hat{e}_U) \cdot \mathbf{k}]\), and background wind speed. Transfer functions for wind divergence \((\hat{A}_3)\) are estimated from \( \hat{A} \) as

\[\hat{A}_3 = (-i)[\hat{e}_U \cdot \mathbf{k} \hat{A}_1 + (\hat{e}_3 \times \hat{e}_U) \cdot \mathbf{k} \hat{A}_2], \tag{A6}\]

The reality condition implies that

\[\hat{A}(k_0, U_0) = \hat{A}^*(-k_0, U_0), \tag{A7}\]

where the star indicates a complex conjugate. For the real part, the first and third quadrants should have the same values, and so do the second and fourth quadrants. The same is the case for the imaginary part, but each pair has the opposite sign.

\[c. \text{Estimation of transfer functions using least squares fitting}\]

A transfer function value at an arbitrary wavenumber and background wind speed is estimated from base transfer function values via a weighted average, namely,

\[\hat{A}[\hat{e}_U \cdot \mathbf{k}, (\hat{e}_3 \times \hat{e}_U) \cdot \mathbf{k}, U] = \sum_{k_0, U_0} W_{e_U, k_0, k_0, U_0} \hat{A}(k_0, U_0), \tag{A8}\]

where \( k_0 \) and \( U_0 \) signify base wavenumbers and base background wind speeds, respectively, and

\[W_{e_U, k_0, k_0, U_0} = W_0 \max \left(1 - \sqrt{\frac{|\hat{e}_U \cdot \mathbf{k}(\hat{e}_3 \times \hat{e}_U) \cdot \mathbf{k}|}{\lambda^2} - \frac{(U - U_0)^2}{\Delta^2}}, 0 \right), \tag{A9}\]

\( \lambda \) and \( \Delta \) are resolution of base wavenumbers and background wind speeds, respectively. The value of \( W_0 \) is selected such that

\[\sum_{k_0, U_0} W_{e_U, k_0, k_0, U_0} = 1. \tag{A10}\]

Transfer functions are obtained using least squares fitting by minimizing a cost function \([\mathbf{R} = (R_1, R_2)]\). For example, \( R_1 \) is expressed as

\[R_1 = \frac{1}{N_x} \sum_x \delta u_1^2 + \frac{\rho^2 T_{RMS}^2}{N_0} \sum_{k_0, U_0} \hat{A}_1 \hat{A}_1^*, \tag{A11}\]

where \( x \) signifies the position vector that covers a target domain, \( N_x \) the number of observations used for the fitting, \( N_0 \) the number of unknown values of \( \Lambda_1 \), \( \rho \) the regularization factor, and \( T_{RMS} \) the root-mean square of high-pass-filtered SST (Fig. 6). To obtain smooth transfer functions, we use \( \rho \) of unity. When estimating “midlatitude” transfer functions, \( x \) in Eq. (A11) covers all the Gulf Stream, Kuroshio, and Agulhas Retrofection domains after the \( f \) scaling that is described just below. The midlatitude transfer functions may be weighted by the number of observation points for each region, which is negligible in the present study.

To facilitate the comparison between the Northern and Southern Hemispheres, crosswind components and crosswind axes are scaled by a factor of \([f/f_0]\), where \( f \) signifies typical Coriolis parameter for each target domain. Thus, \([f/f_0] = 1 ( -1 )\) for the Northern (Southern) Hemisphere. The scaled crosswind components and crosswind axes point toward left (right) relative to background wind direction in the Northern (Southern) Hemisphere.

For every target point within the black lines in Fig. 2, Fourier transform of SST are estimated for a rectangular box with \( 8^\circ \times 8^\circ \) size centered at the target point, and then observed wind components at the target point are fitted using Eq. (A5). The base background wind speeds are shown in Fig. 3. For the base wavenumbers of the transfer...
functions, we use $0.11 \times 10^{-5}$ cycle m$^{-1}$ intervals for both lee and port wavenumbers as the wavenumber resolution, which correspond to meridional wavenumber intervals for SST Fourier transform. Thus, the maximum wavelength resolved is approximately 910 km cycle$^{-1}$, which is still partly the mesoscale range (e.g., Laurindo et al. 2019). Since spatial averages over the 8° boxes are not removed for Fourier transform and the high-pass filter cannot completely remove large-scale signals, the correlation between surface winds and SST on the horizontal scale larger than 910 km cycle$^{-1}$, if any, is reflected at the origin of the transfer functions. The base wavenumbers with magnitude larger than $0.4 \times 10^{-5}$ cycle m$^{-1}$ are excluded from the calculation.

Transfer functions for sea level pressure ($\tilde{A}_P$) and air temperature at 2-m height ($\tilde{T}_{2m}$) are estimated in the same way as in surface winds. Transfer functions for the Laplacian of sea level pressure ($\tilde{\nabla}^2 P$) is estimated as

$$\tilde{\nabla}^2 P = -k_0^2 \tilde{A}_P. \quad (A12)$$

By subtracting the surface air temperature ($T_{2m}$) from SST in the form of inverse Fourier transform as in Eqs. (A5) and (A4), we obtain

$$T - T_{2m} = (2\pi)^{-2} \int dk \left(1 - \tilde{T}_{2m}\right) \tilde{T}(k)e^{-ikx} + \delta(T - T_{2m}).$$

(A13)

Thus, transfer functions for the difference in surface air temperature and SST ($\tilde{A}_{AT}$) should be equal to $\left(1 - \tilde{T}_{2m}\right)$. However, this is not the case with the current formulation because the regularization inhibits large amplitude without referring to the physical consistency. We are exploring other methods to avoid the inconsistency. For now, $\tilde{A}_{AT}$ are estimated by applying the least squares fitting to the temperature difference estimated in the physical domain beforehand.

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