Spectral Descriptions of Submesoscale Surface Circulation in a Coastal Region

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Abstract Submesoscale coastal surface currents at hourly and O(1) km resolutions, obtained from an array of high-frequency radars in a coastal region off the east coast of Korea over a period of 1 year (2013), are described in the frequency and wavenumber domains. The low-frequency surface currents exhibit more consistent variability with the regional geostrophic currents in summer than those in winter because of the relatively weak wind conditions in summer. The clockwise near-inertial surface currents show onshore phase propagation and decreasing amplitudes. The kinetic energy spectra of the surface currents in the wavenumber domain (k) become steeper, increasing from a slope of k^{-5/3} at a length scale of approximately 10 km to slopes between k^{-2} and k^{-3} at a length scale of 2 km. These kinetic energy spectra exhibit anisotropy and weak seasonality with an injection scale of O(10) km, consistent with dominant length scales of the regional submesoscale eddies and the zero-crossing wavenumber in the estimated kinetic energy fluxes. We suggest possible mechanisms to explain the above findings. Submesoscale processes in this region are primarily initiated by surface frontogenesis caused by regional mesoscale currents, including low-frequency circulations and topographically linked shear currents, then maintained by baroclinic instability in the mixed layer with moderate seasonality.

1. Introduction

Oceonic submesoscale processes, which are frequently observed as filaments, fronts, and eddies, are characterized by both Rossby numbers and Richardson numbers of O(1) and have horizontal spatial scales that are smaller than the first baroclinic deformation radius (e.g., Callies et al., 2015; Thomas et al., 2008). The dynamics of these submesoscale processes has been primarily explained in terms of mesoscale eddy-derived surface frontogenesis, baroclinic instabilities in the mixed layer, topographically linked vortices, and turbulent thermal winds (e.g., Callies et al., 2015; McWilliams, 2016). Observations of submesoscale processes are sparse because traditional in situ ocean measurements are limited in their ability to resolve the detailed horizontal and vertical structures of these processes (e.g., Buckingham et al., 2016; Soh & Kim, 2018). Although technological advances have resulted in increases in the density of sampling rates over time and space and the development of the Lagrangian sampling strategy, which involves following and crossing the target phenomena, these observations have mainly focused on specific submesoscale events and phenomena and are available for a limited number of realizations (e.g., Baschek & Molemaker, 2010; Callies et al., 2015; D’Asaro et al., 2011). In contrast, operational and concurrent Eulerian observations using remote sensing instruments can thoroughly investigate submesoscale processes via statistical analyses using available data such as high-frequency radar (HFR)-derived surface currents (e.g., Kim et al., 2011; Kirincich, 2016; Lai et al., 2017) and geostationary ocean color imagery (GOCI)-derived surface concentration maps of chlorophyll, total suspended solids, and colored dissolved organic matter (e.g., Choi et al., 2012b), which are available at hourly and kilometer-scale resolutions and are abundant relative to classic in situ measurements.

Coastal circulation, which is a part of air-sea-land interactions and geophysical boundary layer flows, is typically characterized by complex oceanic responses to wind stress, tides, subinertial low-frequency forces, heat fluxes, and interactions among these oceanic responses, as well as near-inertial currents and the currents induced by bathymetric and shoreline boundaries (e.g., Allen, 1980; Brink, 1991; Lentz & Frewings, 2012; Winant, 1980). Low-frequency coastal circulation, including geostrophic currents, coastally trapped waves, and intermittent interannual climate signals, is related to regional atmospheric variability (e.g., Brink, 1991; Halliwell...
The spectral decay slopes of the one-dimensional wavenumber domain ($k$) energy spectra of dynamic variables (e.g., currents or density) have been used to delineate geostrophic turbulence theories, including the quasi-geostrophic (QG), surface QG (sQG), finite-depth sQG (fsQG), and semi-QG (SG) theories (e.g., Armi & Flament, 1985; Gage, 1979; Kraichnan, 1967; Lesieur & Sadourny, 1981; Soh & Kim, 2018; Vallis, 2006). The decay slopes of these energy spectra become steeper or flatter depending on the pathways of energy and enstrophy. In other words, the decay slopes are modified at the injection scale, where forward enstrophy cascades occur toward small scales and inverse energy cascades occur toward large scales (e.g., Ferrari & Wunsch, 2009; Soh & Kim, 2018; Vallis, 2006). Under the QG theory, the energy spectra of currents have slopes of $-5/3$ for inverse cascades and $-3$ for forward cascades (e.g., Charney, 1971; Gage, 1979; Vallis, 2006). The spectral decay slopes of currents in the sQG theory are $-1$ for inverse cascades and $-5/3$ for forward cascades (e.g., Tulloch & Smith, 2006). The fsQG theory has the QG-like property of a $-3$ slope at large scales and the sQG-like feature of a $-5/3$ slope at small scales (e.g., Tulloch & Smith, 2006). The energy spectra of currents in the SG theory follow a slope of $-9/5$ at small scales (e.g., Andrews & Hoskins, 1978). The energy spectra of surface currents under rough bathymetric effects have a slope of $-2.5$ (e.g., Nikurashin et al., 2013) (see Table 1 for more details).

The primary motivation of this paper is to examine the variance of the observed coastal surface currents with hourly and O(1) km resolutions in the frequency and wavenumber domains to evaluate (1) the consistency between the coastal surface currents understood as the responses to primary driving forces (e.g., subinertial low-frequency forcing, wind stress, and tides) and concurrent observations (e.g., altimeter-derived geostrophic currents, historical hydrographic surveys for temperature and salinity profiles, and high-resolution surface maps of chlorophyll concentrations) and (2) the feasibility of employing coastal surface currents as a resource in studies of submesoscale processes. Prior to the use of HFR-derived surface currents in scientific studies, a certain level of data quality is required. Here an extensive and systematic analysis to improve the quality and fidelity of the HFR-derived surface currents, particularly radial velocity maps, using various metrics is presented (Appendix A).

To carry out this study, we chose a coastal region of Imwon, which is located along the east coast of Korea (Figures 1a and 1b). Here (1) submesoscale eddies and fronts are frequently observed in the GOCl-derived chlorophyll concentrations and HFR-derived surface currents (Figures 1c and 1d) and (2) concurrent surface and subsurface observations are available at a resolution at which submesoscale processes can be resolved (Figures 1c–1f). In this area, the coastal circulation is influenced by the variability of the regional mesoscale boundary currents, specifically the North Korea Cold Current (NKCC) and the East Korea Warm Current (EKWC) (see Choi et al., 2012a for more details) (Figure 1b). The confluence of these two regional boundary currents, which have opposite flow directions and contrasting water properties, leads to the generation of a regional front (i.e., the subpolar front (SPF) in the East/Japan Sea (EJS); Yoshikawa et al., 2012). Studies of the coastal circulation in this region have employed observations using satellite altimetry (e.g., Choi et al., 2012a; Lee et al., 2009), hydrographic surveys (e.g., Shin et al., 1996), and in situ current profiles (e.g., Cho et al., 2014; Lee & Chang, 2014; Park et al., 2016). In particular, onshore propagating (internal) near-inertial waves have been reported from various in situ surface and subsurface observations made using current meters, echo sounders, and synthetic aperture radar (e.g., Kim et al., 2001, 2005a, 2005b; Lie, 1988).

We begin with a review of the individual observations obtained within the study domain (section 2). Then, the primary variance of the surface currents in the frequency domain (e.g., low-frequency and near-inertial frequency bands) and the wavenumber domain energy spectra of the surface currents are presented

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**Table 1**

<table>
<thead>
<tr>
<th>Energy spectra</th>
<th>Inverse cascades</th>
<th>Forward cascades</th>
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<tbody>
<tr>
<td>QG</td>
<td>$k^{-5/3}$</td>
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<tr>
<td>sQG</td>
<td>$k^{-1}$</td>
<td>$k^{-3}$</td>
</tr>
<tr>
<td>fsQG</td>
<td>$k^{-5/3}$</td>
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The spectral decay slopes of the one-dimensional wavenumber domain ($k$) energy spectra of currents ($S(k)$) are listed in terms of geostrophic turbulence theories (Quasi-Geostrophic (QG), Surface QG (sQG), and Finite-Depth sQG (fsQG) Theories) and the directions of the energy pathways (Inverse Cascades, Forward Cascades, and Surface Dissipation) (e.g., Tulloch & Smith, 2006; Vallis, 2006). Coastal currents driven by barotropic and baroclinic tides have been identified by narrow peaks (e.g., tidal lines) and cuspy peaks (e.g., peaks exhibiting similar widths in the low-frequency variance), respectively, in their energy spectra (e.g., Hendershott & Munk, 1970; Munk et al., 1965). In contrast, coastal diurnal currents are primarily associated with diurnal tides, heat fluxes, and wind stress (e.g., land-sea breezes due to the daily development of the marine boundary layer) (e.g., Kim & Crawford, 2014; O’Brien, 1974; Price et al., 1986; Wexler, 1946). The near-inertial currents and their coastal inhibition have been ubiquitously observed in coastal regions (e.g., D’Asaro, 1985; Davies & Xing, 2003; Kim et al., 2015; Kim & Kosro, 2013; Shearman, 2005).
(sections 3.1 and 3.2). A discussion of the results and conclusions are included in sections 4 and 5, respectively. Note that the power spectral density is referred to as a spectrum in the remainder of the paper, and $x$ and $y$ directions indicate cross-shore and alongshore directions, respectively.
2. Data

2.1. Coastal Surface Currents

Hourly averaged observations of surface currents were obtained off the coast of Imwon, Republic of Korea, on a grid with a 1 km spatial resolution over a period of 1 year (2013) using two phased-array HFRs (WEllen RAdar (WERA) systems; Gurgel et al., 1999a, 1999b)–Imwon North (IMWN; R1) and Imwon South (IMWS; R2) (Figure 1c). Although WERA-derived radial velocities are typically reported on a Cartesian coordinate grid, which enables users to conveniently map vector currents (e.g., Voulgaris et al., 2011), we internally modified the computing scripts to report the radial velocities on the polar coordinate grids of individual HFRs. This allows us to investigate the quality of the radial velocity data prior to mapping and to apply optimal interpolation (OI) to directly estimate the vector currents as well as the kinematic and dynamic quantities associated with surface currents (e.g., stream function, velocity potential, divergence, vorticity, and strain rates) without any intermediate steps, thus minimizing the amplification of noise in the estimates. For instance, divergence and vorticity are quantities that represent the spatial derivatives of vector currents, and the spatial derivatives of noisy data fields can contain significantly amplified noise. For the OI method in this paper, an exponential correlation function with an isotropic decorrelation length scale of 1.5 km and a search radius of 3 km under the cutoff correlation of 0.1 is applied to minimize spatial smoothing on the radial velocity maps with resolutions longer than 1 km (Kim et al., 2008, 2007; Soh et al., 2018).

The spatial data availability of the radial velocities at each radar (IMWN and IMWS) and of the vector currents over a period of 1 year (2013) is shown in Figures 2a–2c, respectively. The spatial data availability at a single location represents the number of observations normalized by the total number of time stamps (i.e., 8,544 realizations are performed in 1 year) (Kim, 2015). The effective study area is marked with a gray curve in Figures 1c and 2c, based on the quality assurance and quality control (QAQC) of the radial velocities (Appendix A). The temporal data availability of the radial velocities at each radar (IMWN and IMWS) and of the vector currents for the same time period is shown in Figures 2d and 2e, respectively. Similarly, the temporal data availability at a single time represents the number of observations normalized by the maximum of the expected number of radial solutions or vector solutions (Kim, 2015). Based on the definitions of spatial and temporal data availability, a value of zero indicates that no data are available and a value of one indicates that complete data are available over the given domain or at a given time.

Both radars were operated with 24.525 MHz as the central frequency under slightly different bandwidth configurations. The radar at IMWN was operated with a fixed bandwidth (150 kHz), and the radar at IMWS employed three bandwidths (150, 125, and 100 kHz) depending on the sea states and the optimal signal-to-noise ratios of the Bragg backscattered signals. These bandwidths yielded the different range spacings ($\Delta s = 1, 1.2, \text{and} 1.5 \text{km}$) and resulted in the generation of radial velocity maps on three different polar grids.
The maps of the spatial data availability obtained from the three different range spacings at IMWS have similar spatial patterns in terms of their maximum ranges and azimuthal widths. The radial velocities obtained at IMWN are more intermittent than those sampled at IMWS (Figure 2d).

The outliers in the radial velocity maps can be identified using the residuals obtained from the inverse methods (e.g., unweighted least squares fitting or OI), which are used to extract a vector current map from multiple radial velocity maps (e.g., Kim et al., 2008). However, intermittent or spatially uniform outliers may not be detectable. Thus, the outliers are isolated from the synoptic-scale regional surface oceanic responses (e.g., Lee & Chang, 2014) using (1) the finite time differences of the time series of the radial velocities and vector currents, i.e., the acceleration term of the flow fields, and (2) the running means and standard deviations of individual time series calculated using a 5 day time window, following the procedures used in earlier studies (e.g., Kim et al., 2007). The outliers identified using these procedures are not included in the following analysis.

### 2.2. Coastal Winds

Hourly wind data recorded over a period of 1 year (2013) at a shore station (W1) and an offshore wind buoy (W2) were obtained from the Korea Meteorological Agency (KMA) (Figures 1c and 1d). Both nearshore and offshore winds have the dominant variance and coherence in the subdiurnal frequency band $|\sigma| \leq 1$ cycles per day (cpd) (Figures 3a and 3b). The linearly increasing phase of the coherence between the two wind datasets in the clockwise subdiurnal frequency band indicates that the offshore wind leads the nearshore wind with nearly constant time lags (Figure 3c). Note that the wind data obtained at the W1 station and W2 buoy are missing 33% and 1% of their individual time records, respectively. Their spectral estimates are

![Figure 2](image-url):

(a–c) Spatial data availability (a dimensionless quantity) of the HFR-derived hourly radial velocities and vector currents for a period of 1 year (2013). (a) IMWN (R1), (b) IMWS (R2), and (c) vector currents. The spatial data availability at a single location represents the number of observations normalized by the total number of time stamps (8,544 realizations are performed in 1 year). A gray contour in Figure 2c indicates an effective area having the expected correlations and standard deviations, obtained from a QAQC evaluation on the paired radial velocities (see Appendix A). (d and e) Temporal data availability (a dimensionless quantity) of the HFR-derived hourly (d) radial velocities at IMWN and IMWS, and (e) vector currents for a period of 1 year (2013). The temporal data availability at a single time represents the number of observations normalized by the maximum of the expected number of radial solutions or vector solutions. Based on the definitions of the temporal and spatial data availability, zero indicates that no data are available and one indicates that complete data are available over the given domain or at a given time. The temporal data availability lower than 0.2 is marked as crosses (less than 3% of total number of realizations) to minimize overlapping time series. Note that Figures 2b and 2d are obtained from the radial velocity maps at a range spacing of 1 km.
based on the slow Fourier transform, which is a least squares fit that uses all orthogonal basis functions (frequencies) and appropriately treats temporally missing data (e.g., Soh & Kim, 2018).

The temporal variability of the nearshore and offshore winds is presented with the time series of the probability density functions (PDFs) of the wind speeds and directions at the W1 station and W2 buoy (Figure 4). Individual PDFs are estimated using the data obtained within a 10-day time window. The wind speeds at both stations become reduced in summer and enhanced in winter, even though there are different degrees of magnitudes in the seasonal differences of the wind speed (Figures 4a and 4b). The dominant wind directions vary seasonally southeastward (winter), northeastward (spring), northwestward (summer), to southwestward (fall), with minor directional deviations and differences between nearshore and offshore winds (Figures 4c and 4d). The down-front winds with respect to the regional SPF (Figure 1b) are set up in spring and early summer (i.e., the rectangular black boxes marked in Figures 4c and 4d).

2.3. Temperature and Salinity Profiles
Temperature and salinity data were obtained from conductivity-temperature-depth (CTD) casts collected at the C0, C4, C5, and C6 stations along the hydrographic survey Line 104 of the National Fisheries Research and Development Institute (NFRDI) during the recent 20 year period extending from 1995 to 2014 (Figure 1c). These data are used to identify the seasonal stratification and seasonal time windows in this region. The temperature records covering this 20 year period, which have an approximately bimonthly temporal resolution, are regressed using the temporal mean, seasonal frequency, and its five superharmonic frequencies as basis functions. Figures 5a and 5b show the raw surface temperature data, the reconstructed time series on a monthly time axis, and the residuals at the C5 station (Figure 1c), respectively. The regressed time series (the black curve in Figure 5a) explains approximately 88% of the variance of the raw data (the red dots in Figure 5a). Figures 5c, 5e, 5g, and 5i show the vertical profiles of temperature when the surface temperatures reach their maximum (25 August 2013) and minimum (24 February 2013) values at the four stations. Similarly, the time series of the reconstructed temperature profiles on a time axis with a resolution of 3 days over a period of 1 year (2013) are shown in Figures 5d, 5f, 5h, and 5j.

In this paper, the isopycnal depths of 25.5 and 26.5 kg m$^{-3}$ are chosen to present the variability of vertical stratification as the regional mixed layer depth (MLD) since it is difficult to identify the MLD using vertical temperature and density gradients (e.g., de Boyer Montégut et al., 2004; Holte & Talley, 2009; Kara et al., 2000) due to the negligible vertical variations in density, particularly in winter.

2.4. Satellite-Derived Geostrophic Currents and Sea Surface Temperature
Daily data on AVISO geostrophic currents with a resolution of 0.25° over a period of 1 year (2013) (Le Traon et al., 1998; see the Acknowledgements for the acronym) are analyzed to examine: (1) the large-scale context of the regional low-frequency circulation and mesoscale variability (e.g., eddies and fronts) and (2) the geostrophic contribution to the coastal surface currents and details of submesoscale surface currents (Figure 1e).

Daily OSTIA UK Met Office (UKMO) sea surface temperature (SST) maps at a grid resolution of 0.05°, based on multiple satellite sensors and in situ data from drifting and moored buoys, are analyzed (e.g., Stark et al., 2007) (Figure 1f). Although the Jet Propulsion Laboratory (JPL) SST product covers this region, the OSTIA UKMO product was used in the analyses described below because the JPL SST product does not clearly
represent the spatial and temporal variability of the regional SPF, which is a dominant large-scale feature of the circulation in the area of interest.

Two orthogonal grid lines are drawn normal (PN–PS line) and parallel (PW–PE line) to the regional SPF to examine the regional mesoscale variability in the geostrophic currents, the seasonal anomalies in SST, and their spatial gradients (Figures 1e and 1f).

3. Results

3.1. Variance in the Frequency Domain

In this section, we estimate the frequency domain energy spectra of the surface currents (section 3.1.1) and interpret the variances in the primary frequency bands in terms of their temporal variability and spatial structure (sections 3.1.2 and 3.1.3).

3.1.1. Estimates of the Energy Spectrum

The frequency domain ($\gamma$) energy spectrum $[S(\gamma)]$ is estimated with using the spatial average of the energy spectra $[S(x, y, \gamma)]$ of the surface currents ($u=U+iV$) at individual locations ($x, y$):

$$S(\gamma) = \langle S(x, y, \gamma) \rangle_{x,y},$$

where

$$S(x, y, \gamma) = \left| \int_{-\infty}^{\infty} u(x, y, t)e^{-i\gamma t}dt \right|^2$$

and $\langle \cdot \rangle_{\cdot}$ denotes the (ensemble) average with respect to the subscript.

Significant variance in the surface currents appears in two frequency bands, specifically, the nonzero low-frequency band ($0 < |\gamma| < 0.2$ cpd) for both rotations (clockwise and counter-clockwise) and the clockwise near-inertial frequency band ($-1.4 \leq \gamma \leq -1.2$ cpd) (Figure 6a). Although the variance in the surface currents at the clockwise diurnal frequency ($\gamma=-1$ cpd) and in the clockwise semidiurnal frequency band (centered by the $M_2$ frequency, $\gamma=-1.932$ cpd) is above the noise level, we do not include these currents in the
following analysis, considering the relative insignificance of their magnitudes. The OI-mapped normalized vorticity or Rossby number ($R_o = \zeta / f$, where $\zeta = \partial v / \partial x - \partial u / \partial y$ and $f$ denote the vertical component of the relative vorticity and local Coriolis frequency, respectively) follows a nearly Gaussian distribution within $|R_o| \leq 3$ (Figure 6b) with an asymmetric shape that the distribution becomes positively slanted under a range of $0 \leq |R_o| \leq 0.5$ and negatively slanted under a range of $0.5 \leq |R_o| \leq 1$. Note that the Rossby numbers of

Figure 5. (a and b) An example of (a) the bimonthly surface temperature records at CS station for recent 20 years (1995–2014; red dots), the reconstructed time series on a monthly time axis (black), and (b) residuals. (c–j) Seasonal temperature profiles and time series of the reconstructed temperature profiles (°C) on a 3 daily time axis at C0, C4, C5, and C6 stations along the hydrographic survey Line 104 (Figure 1c). (c), (e), (g), and (i): Seasonal temperature profiles. Paler and darker colors indicate summer (s) and winter (w) profiles, respectively. (d), (f), (h), and (j): Time series of the temperature profiles on a 3 daily time axis for a period of 1 year (2013), reconstructed from a regression analysis using basis functions of a temporal mean, seasonal frequency, and five harmonic frequencies on the approximately bimonthly sampled temperature records for recent 20 years (1995–2014) (section 2.3). The isopycnals are marked with a density interval of 0.5 kg m$^{-3}$, and the density anomalies ($\rho' = \rho - 1.000$ kg m$^{-3}$) of 25.5 and 26.5 kg m$^{-3}$ are highlighted with red.
greater than 3 represent less than 1% of the entire dataset (2013), and they are thus excluded from the following analysis.

The variability ($\xi$; equation (3)) of the hourly surface currents in the primary frequency bands is presented as a map of their magnitude, which is the square root of the area under the frequency domain energy spectrum ($S(\sigma)$; equation (1)) within the frequency band of interest ($\sigma_1/C_{20}$) (e.g., Priestley, 1981) (Figures 7b and 7c):

$$
\xi = \left[ \sum_{\sigma} S(\sigma) \Delta \sigma \right]^{\frac{1}{2}}.
$$

(3)

3.1.2. Temporal Mean and Low-Frequency Variability

The temporal means ($\sigma = 0$ cpd) and standard deviations of the surface currents over a period of 1 year (2013) show spatially uniform northward flows along with significant offshore variability (Figure 7a), which are part of the dominant northward EKWC throughout the year (e.g., Choi et al., 2012a; Lee & Chang, 2014).

The low-frequency surface currents are enhanced offshore and reduced nearshore (Figure 7b). This pattern is similar to the spatial structure of the standard deviations of the surface currents (Figure 7a), which is qualitatively consistent with the frictional effects of the coastal boundary on the subinertial surface currents (Figure 7b). The reduced variance of the low-frequency surface currents at the eastern boundary of the study domain may be caused by the loss of signal strength at far ranges.

The satellite-derived geostrophic currents near the coast may be contaminated due to the presence of land boundaries (e.g., Saraceno et al., 2008) and may have missing information below their temporal and spatial scales (e.g., $O(1)$ week and $O(100)$ km, Stammer, 1997). Despite this contamination, the low-frequency surface currents and the altimeter-derived geostrophic currents are compared to evaluate the consistency in the time and frequency domains and to highlight detailed information in the submesoscale observations.

As a time domain analysis, the two-daily averaged components of the surface currents over the entire domain ($\mathbf{u}_{\text{SFC}}^{\text{geo}}$; darker colors) are overlaid on the envelope of standard deviations of the components of the hourly surface currents over the entire domain ($\mathbf{u}_{\text{SFC}}^{\text{geo}}$; paler colors) to represent the low-frequency variability of the surface currents in the study domain and to compare them with the daily geostrophic currents, both at G1 ($\mathbf{u}_{\text{geo}}$) and averaged over three locations (G1 to G3; $\mathbf{u}_{\text{geo}}^{\text{avg}}$) (Figure 8; see Figure 1c for the sampling locations). The 2 day long time window is chosen as a conservative frequency limit to represent the low-frequency variance in the surface currents ($\sigma < 0.5$ cpd).

As a frequency domain analysis, as the estimates of coherence between two time series require that they have the same record lengths and sampling intervals, the surface currents are averaged daily ($\mathbf{u}_{\text{SFC}}^{\text{geo}}$). The
magnitudes of the coherence (\( \zeta \) in equation (4)) between the daily averaged surface currents at the individual grid points (\( u_{dd, SFC} \)) and the daily geostrophic currents (\( u_{dd, GEO} \)) at the location of G1 are presented as PDFs in individual frequency bins (Figure 9a):

**Figure 7.** (a) Temporal means (\( \sigma = 0 \) cpd) and standard deviations of surface currents (cm s\(^{-1}\)) are presented with the black arrows and color-coded patches, respectively. The temporal mean of surface currents is plotted at every 2 km to minimize overlapping of the arrows. (b and c) Magnitudes (cm s\(^{-1}\); \( \zeta \) in equation (3)) of variability of the hourly surface currents for a period of 1 year (2013) (b) in the nonzero low-frequency band (0 < \( |\sigma| \leq 0.2 \) cpd) on both rotations and (c) in the clockwise near-inertial frequency band (\(-1.4 \leq \sigma \leq -1.2 \) cpd). The bottom bathymetry is contoured at 50, 100, 200, 300, and 400 m.

**Figure 8.** A comparison of the time series of the two-independent current observations. The two-daily averaged components of the surface currents over the entire domain (\( u_{dd, SFC} \); darker colors), which are overlaid on the standard deviations of components of the hourly surface currents over the entire domain (\( u_{sh, SFC} \); paler colors), are compared with the daily AVISO geostrophic currents at G1 in Figure 1c (\( u_{dd, GEO} \); black) and averaged over three locations of G1 to G3 in Figure 1c (\( u_{dd, GEO} \); gray). (a) Zonal component (\( u; \) cm s\(^{-1}\)). (b) Meridional component (\( v; \) cm s\(^{-1}\)).
The altimeter-derived geostrophic currents mostly lie within the envelope given by the standard deviations of the HFR-derived subinertial surface currents with more consistent variability in both current components between late June and July (summer) and moderate differences between them in October and December (Figure 8) because of the small contribution of ageostrophic components in summer under the relatively weak wind conditions (Figures 4a and 4b). In particular, the meridional component of the HFR-derived subinertial surface currents has steeper reversals in the early summer than that of the altimeter-derived geostrophic currents (Figure 8b), which has been referred to as the “summer current reversal” in this region and is consistent with the previously observed regional coastal currents produced by the regional mesoscale boundary currents (e.g., Lee and Chang, 2014), alongshore buoyancy gradients (e.g., Park et al., 2016), and subinertial temperature variability (e.g., Cho et al., 2014). The magnitudes of coherence in all three subinertial frequency bands ($|\sigma_1| \leq 0.0192$, $0 < |\sigma_2| \leq 0.1$, and $0 < |\sigma_3| \leq 0.25$) are presented as maps to evaluate the spatial consistency between the two low-frequency currents (Figures 9b–9d).

3.1.3. Near-Inertial Variability

The near-inertial motions appear ubiquitously as part of the resonant ocean responses to broadband wind forcing (e.g., Kim and Kosro, 2013; Kunze, 1985a), and the near-inertial (internal) waves have horizontally dominant motions in terms of their wavelengths and amplitudes compared with the (internal) waves near the buoyancy frequency ($N$) as the distinct characteristics of the internal waves are determined by the frequency within the lower and upper bounds ($N^2 \leq \omega^2 \leq N^2$). When the near-inertial (internal) waves are assumed in a propagating mode (e.g., Broutman et al., 2004; Jones, 1969; Kim et al., 2015), the slope (c) of the propagating ray in the cross-shore direction is given as,

$$c = \frac{\omega^2 - \omega_{\text{crit}}^2}{N^2 - \sigma^2}$$

where $\sigma$ denotes the frequency of the signals of interest. The slope of the ray is the angle to the horizontal along which the characteristics of the internal wave travel (e.g., Cacchione et al., 2002).
The frequency corresponding to the peak near-inertial variance can be shifted with the amount of background vorticity (e.g., Kim & Kosro, 2013; Kunze, 1985a, 1985b). A subinertial shift in the near-inertial variance may transform the inertial motions into an evanescent mode with trapped waves, while a superinertial shift can transform them into a propagation mode with propagating waves.

The effective Coriolis frequency \( f_e \) includes the subinertial or superinertial shift relative to the Coriolis frequency due to background vorticity and is equal to

\[
f_e = f_c (1 + \frac{\zeta}{f_c})^2 = f_c (1 + \frac{\zeta}{f_c})^2
\]

(e.g., Whitt & Thomas, 2015), and it has been approximated using a Taylor series:

\[
f_e \approx f_c + \frac{\zeta}{2} f_c \left(1 + \frac{R_o}{2}\right)
\]

(e.g., Kim & Kosro, 2013; Kunze, 1985a, 1985b).

At the transition from the open ocean to the coast, the near-inertial motions are modified from circular (or elliptical under background vorticity) motions to rectilinear motions, and their magnitudes are reduced as a result of the frictional influence of the coastal boundaries (e.g., the ocean bottom and the coastline), which is referred to as “coastal inhibition” (e.g., Kim et al., 2015; Kim & Kosro, 2013).

The observed near-inertial surface currents in this region have dominant clockwise superinertial variance (Figure 6a), which can imply propagating (internal) near-inertial waves within the domain. The spatial coherence of the clockwise near-inertial surface currents between reference locations (shown as a white star for the amplitudes and a black star for the phases; \(|\sigma + f_c| \leq 0.3 \text{ cpd}\) and the other grid points shows the spatially correlated structures and propagating phase patterns (Figure 10). The decorrelation length scales decrease onshore (Figures 10a–10d), and the phases increase southward in the middle of the domain and shoreward (southwestward) near the coast as the propagation direction changes (Figures 10e–10h). This result is consistent with those of previous studies (e.g., Kim et al., 2001, 2005b; Park et al., 2006), particularly the characteristics of the near-inertial currents observed at an in situ buoy located approximately 50 km northwest of the study area (D in Figure 1c).

The superinertial variance of the surface currents can be related to the positive background vorticity, with an amount of 0.1 cpd \((\Delta \sigma = 0.1 \text{ cpd})\) (Figure 6a), which corresponds to Rossby numbers of 0.166 and 0.172,

![Figure 10. Examples of spatial coherence (a dimensionless quantity) of the clockwise near-inertial surface currents \(|\sigma + f_c| \leq 0.3 \text{ cpd}\) between a reference location (a white star for magnitudes and a black star for phases) and the rest of grid points for four different reference locations. (a–d) Magnitudes (dimensionless quantities). (e–h) Phases (°).](image-url)
respectively (equations (6) and (7); $f_c = 1.2064$ cpd at 37.10°N). The positively slanted background vorticity plays a dominant role in modifying the vorticity (Figure 6b), which leads to the superinertially shifted near-inertial variance in the energy spectrum (Figures 6a, 11a, and 11b). The onshore and offshore cross-shore propagations of vorticity are marked with black arrows, based on their organized cross-shore structures (Figures 11a and 11b).

Figure 11.(a) and (b): Time series of (a) the Rossby number ($R_o = \zeta / |f_c|$), i.e., normalized vorticity and (b) the stream function ($\psi$; m$^2$ s$^{-1}$) along the cross-shore line in Figure 1c. Potential propagating directions of vortical features in the cross-shore direction are marked with black arrows. Two gray boxes indicate the time periods to show significant variability of surface vorticity in the cross-shore direction (Figures 11a and 11b) and submesoscale eddies with high vorticity (Figure 14e). 

(c and d) Time series of variance (log$_{10}$ scale; cm$^2$ s$^{-2}$) of the surface currents at the trial frequencies within the clockwise near-inertial frequency band ($|\alpha_{f_c}| \leq 0.5$ cpd; $|\alpha_{f_c}+1| \leq 0.41$), at two sampling locations of A and B in Figure 1c, respectively. The pink and light purple curves indicate the scaled stream function ($\psi/\psi^*$, where $\psi^* = 500$ m$^2$ s$^{-1}$) and the scaled Rossby number ($R_o/R^*$, where $R^*=2$) at each location, respectively. Note that $\psi < 0$ and $R_o < 0$ denote clockwise rotational flows (red) and $\psi > 0$ and $R_o > 0$ denote counter-clockwise rotational flows (blue). The individual estimates are based on a 9 day moving time window along with a 3 day time increment to present smooth variation of amplitudes in time. The vertical axes (y axis) in Figures 11c and 11d are marked for the trial frequency normalized by the local Coriolis frequency ($|\alpha_{f_c}/f_c| \leq 0.41$) on the left and for the scaled normalized vorticity (pink) and stream function (purple) on the right, respectively.
To provide a more detailed explanation of the shift in the near-inertial peak due to background vorticity, we present the time series of the variance in the surface currents at two locations (A and B in Figure 1c) at trial frequencies within the clockwise near-inertial frequency band \( f \approx \frac{\sigma}{f_c} = \frac{1}{1 + R_o} \) or \( f \approx \frac{\sigma}{f_c} = 1 + R_o \), which is estimated using least squares fitting (Figures 11c and 11d) (e.g., Kim & Kosro, 2013). The Rossby number \( (R_o) \) and stream function \( (\psi) \) represent the rotation of the current field as complementary tools (e.g., Kim, 2010). Since the Rossby number, which is a quantity of the spatial derivatives of surface vector currents, may contain more observational noise, the sign of the Rossby number may not be sufficient to determine the rotational direction of the flow. Thus, the stream function, which is a quantity of the spatial integration of surface vector currents, is used to verify the sign of rotation. Note that \( \psi > 0 \) and \( R_o > 0 \) indicate clockwise rotational flows (red), whereas \( \psi < 0 \) and \( R_o < 0 \) indicate counter-clockwise rotational flows (blue). The time series of the scaled stream function \( (\psi/\psi_p, \text{where } \psi_p = 500 \text{ m}^2 \text{ s}^{-1}) \) and the scaled Rossby number \( (R_o/\psi_p, \text{where } \psi_p = 2) \) at the two locations (A and B in Figure 1c) are closely correlated with the enhanced amplitudes at the trial frequencies (Figures 11c and 11d). The amount of frequency \( (\delta f; \delta f = f_e - f_c) \), by which the clockwise near-inertial peak is shifted and the contribution of background vorticity \( (\delta f = f_e - f_c - 1) \), as estimated using equations (6) and (7) at two locations (A and B), are compared (Figures 12a and 12b). Note that \( \delta f \) and \( \delta \sigma \) are dimensionless quantities. Although both estimates do not provide a one-to-one correspondence (a gray diagonal line), they have similar orders of magnitude. The estimates using equation (6) are more sensitive to high normalized vorticity than those using equation (7) (Figure 12a). In addition, the submesoscale vorticity propagates in the cross-shore direction (black arrows in Figures 11a and 11b), Note that the propagating features are visually identified with their enhanced and organized patterns and have speeds ranging from 0.7 to 1.4 km d \(^{-1}\) in the cross-shore direction. Note that the stream function and Rossby number are scaled using constant values so that they can be shown within a comparable range. Two different near-inertial frequency bands are chosen for a conservative analysis by (1) searching the shifted peaks in a broad frequency band \( (|\sigma + f_e| \leq 0.5 \text{ cpd}) \) to have a better chance to find them and (2) calculating the near-inertial variance in a narrow frequency band \( (|\sigma + f_e| \leq 0.3 \text{ cpd}) \) to minimize the extra variance included in the variance estimates.

The magnitudes of the clockwise near-inertial surface currents are reduced near the coast and enhanced offshore (Figure 7c); this spatial pattern is suggestive of coastal inhibition. The variance of the counter-clockwise near-inertial surface currents is not negligible (Figure 6a); this variance may be consistent with...
the resonant response of the near-inertial motions under sheared geostrophic flows, rather than the rectilinear motions near the coast because of the significant differences in the order of the near-inertial variance in both rotations (e.g., Kim & Kosro, 2013; Whitt & Thomas, 2015). Thus, the near-inertial surface currents appear as elliptical motions instead of purely circular motions.

The observations of the cross-shore propagation of (1) the submesoscale vorticity (Figures 11a and 11b) and (2) the phase of the clockwise near-inertial motions (Figure 10) may be unique in modern studies using remote sensing technology and may stimulate additional studies in physics and biology (e.g., Mahadevan & Tandon, 2006; Omand et al., 2015; Taylor & Ferrari, 2011). For instance, the onshore and offshore propagating submesoscale vortical flows are closely related to near-coast chlorophyll blooms (e.g., Lee & Kim, 2018). In a similar context, the vertical secondary circulation associated with submesoscale vortical flows has been reported in a coastal region (e.g., Kim, 2010).

### 3.2. Variance in the Wavenumber Domain

In this section, we estimate the one-dimensional wavenumber domain energy spectra of the surface currents and their spectral decay slopes (sections 3.2.1 and 3.2.2) and interpret them in the context of regional submesoscale and mesoscale variability, including mixed layer depths (section 3.2.3), submesoscale eddies (section 3.2.4), and mesoscale circulation (section 3.2.5).

#### 3.2.1. Estimates of Energy Spectra

The one-dimensional wavenumber domain energy spectra of the observed radial velocity maps (IMW) and vector current maps (within the black box in the subfigure of Figure 13c) are examined to evaluate the characteristics (e.g., anisotropy and seasonality) and potential drivers (e.g., strain-induced frontogenesis, mixed layer instability, turbulent thermal wind, and topographic wakes) of the submesoscale turbulent flows (Figure 13) (e.g., Callies et al., 2015; McWilliams, 2016). The one-dimensional wavenumber domain energy spectra [\(S_x(k, t)\) and \(S_y(k, t)\)] of the hourly surface vector currents sampled in one direction (i.e., the range, cross-shore, and alongshore directions) are averaged in the other direction (i.e., the azimuthal, alongshore, and cross-shore directions). The spatially averaged one-dimensional wavenumber domain energy spectra \([\hat{S}_x(k)\) and \(\hat{S}_y(k)\)] are then ensemble averaged in time over a period of 1 year (2013) or for individual seasons (summer and winter):

\[
\hat{S}_x(k) = \langle S_x(k, t) \rangle_t, \tag{8}
\]

\[
\hat{S}_y(k) = \langle S_y(k, t) \rangle_t, \tag{9}
\]

where

\[
S_x(k, t) = \left\langle \left\| \int_{-\infty}^{\infty} u(x, y, t) e^{-ikx} dx \right\|^2_y \right\rangle, \tag{10}
\]

\[
S_y(k, t) = \left\langle \left\| \int_{-\infty}^{\infty} u(x, y, t) e^{-iky} dy \right\|^2_x \right\rangle, \tag{11}
\]

\(x\) and \(y\) denote the cross-shore and alongshore directions, respectively.

The 1 year averaged wavenumber domain energy spectra of the observed surface currents decay slopes between \(k^{-2}\) and \(k^{-3}\) at a scale of 2 km (Figures 13a and 13c), with slight differences in the sampling directions (e.g., the cross-shore and alongshore directions). The seasonally averaged wavenumber domain energy spectra show that their decay slopes are nearly identical in two seasons and that their variances are higher in summer than they are in winter (Figure 13d). The variance of the seasonally averaged energy spectra is also dependent on the sampling directions (Figures 13c and 13d). The wavenumber domain energy spectra of the radial velocities at IMWS have similar decay slopes in their averaged estimates over the year and seasons (not shown), and the number of vector current maps participating in the estimates of the spectral decay slopes does not show any seasonal bias (Figure 13g).

#### 3.2.2. Estimates of Spectral Decay Slopes

To perform a more detailed investigation of the temporal variability of the spectral decay slopes, the wavenumber domain energy spectra \([\hat{S}_x(k)\) and \(\hat{S}_y(k)\)] within the individual wavenumber ranges \([0.04 \leq k_x \leq 0.1 \text{ km}^{-1}\) and \(0.15 \leq k_x \leq 0.4 \text{ km}^{-1}\) for \(S_x(k)\); \(0.06 \leq k_y \leq 0.14 \text{ km}^{-1}\) and \(0.15 \leq k_x \leq 0.4 \text{ km}^{-1}\) for \(S_y(k)\)] are regressed using a constant slope (see Figures 13c, 13e, and 13f). The temporal means
Figure 13. (a and b) Wavenumber \( k \) domain energy spectra of the hourly radial velocities sampled in the range direction at IMWN for a period of 1 year (2013) are averaged over the entire year (black) and seasons \( S_r(k_r) \) for summer (red; June, July, and August of 2013) and \( S_r(k_r) \) for winter (blue; January, February, and December of 2013). The wavenumber domain energy spectra of the radial velocities sampled in the range direction are averaged in the azimuthal direction. (c and d) Wavenumber \( k \) domain energy spectra of the hourly vector currents for a period of 1 year (2013) are averaged over the entire year \( S_x(k) \) for cross-shore direction (blue) and \( S_y(k) \) for alongshore direction (red) and seasons \( S_x(k) \) (red) and \( S_y(k) \) (light red) for summer; \( S_x(k) \) (blue) and \( S_y(k) \) (light blue) for winter. The sampling directions of the surface currents are marked with colored arrows in a black box as the subset of study domain in Figure 13c. The wavenumber domain energy spectra of the vector currents sampled in the cross-shore \( x \) and alongshore \( y \) directions are averaged in the alongshore and cross-shore directions, respectively. Gray axillary lines of the spectral decay slopes of \( k^{-2} \), \( k^{-2/3} \), and \( k^{-1} \) are overlaid. (e and f) Spectral decay slopes are estimated from the energy spectra of 6 hourly averaged surface currents using a least squares fit in individual wavenumber ranges \( [0.04 \leq k_1 \leq 0.1 \text{ km}^{-1}] \) (orange) and \( 0.15 \leq k_2 \leq 0.4 \text{ km}^{-1} \) (green) for \( S_{x}(k) \); \( 0.06 \leq k_1 \leq 0.14 \text{ km}^{-1} \) (orange) and \( 0.15 \leq k_2 \leq 0.4 \text{ km}^{-1} \) (green) for \( S_{y}(k) \) (see Figure 13c). Their mean and standard errors within a 10 day long time window are presented with a colored square and vertical line, respectively. The expected spectral decay slopes \( k^{-2} \) (orange) and \( k^{-3} \) (green) are plotted with colored horizontal lines. (e) Spectral decay slopes of \( S_{x}(k) \). (f) Spectral decay slopes of \( S_{y}(k) \). (g) The number of vector current maps participating in the estimates of the spectral decay slopes in Figures 13e and 13f. The bin size is equal to 10 days. (h) Time series of the depth of the density anomaly \( \rho' = 26.5 \text{ kg m}^{-3} \) at C0, C4, C5, and C6 stations (see Figures 5d, 5f, 5h, and 5j).
and standard errors of the estimated spectral decay slopes within a 10
day long time window are presented as a time series with spectral decay slopes of $k^{-5/3}$ (orange) and $k^{-3}$ (green) in the individual wave-
number ranges (Table 1; Figures 13e and 13f). For the estimates of spectral decay slopes, the wavenumber domain energy spectra are
computed from 6 hourly averaged surface vector current maps instead of hourly data because the use of a 6 h long time window can
(1) maintain the diurnal and semidiurnal variability of the surface cur-
cents in this region (the Nyquist frequency is equal to 2 cpd) and (2)
result in improved signal-to-noise ratios (SNRs) within the individual
wavenumber ranges. Note that the chosen low-wavenumber ranges
differ because the wavenumber axes are not identical due to the	nonsquare dimensions of the sampling domain.

The time series of the estimated spectral decay slopes over the 1 year
are close to the decay slopes of $k^{-5/3}$ and $k^{-3}$ for the individual wave-
number ranges of $k_{1x}$, $k_{1y}$, and $k_2$, respectively (Table 1; Figures 13c
and 13d), and they have fluctuations of seasonal superharmonic vari-
ability (Figures 13e and 13f).

### 3.2.3. Variability of Mixed Layer Depths

The MLDs at all four stations (C0, C4, C5, and C6) exhibit nearly consis-
tent variability, having variance at the seasonal and its superharmonic
frequencies, reaching depths of 25–75 m in summer and 50–100 m in
winter, and deepening offshore (Figure 13h). However, the MLD at the
C6 station in summer is inconsistent with those at other stations because
of the influence of a regional mesoscale eddy, the Ulleung Warm Eddy,
which pushes down the thermocline at its center when it passes by near
C6. The regional variability of the MLDs contain both seasonality modu-
lated by seasonal superharmonic frequencies and regional mesoscale cir-
culation (Figure 13h). The temporal variability of the regional MLD is
partly out-of-phase with the variability of the spectral decay slopes in the
wavenumber range of $k_2$ (green in Figures 13e and 13f).

### 3.2.4. Statistics of Identified Submesoscale Eddies

The surface small-scale and submesoscale eddies have been identified
using the geometry of the stream function and vorticity (e.g., Kim, 2010).
The vortical features of the surface current field are captured in the nearly
closed contours of stream functions, which can be detected when the sum
of the exterior angles of the N-segmented piecewise contour is equal to $2\pi$,
i.e., the winding angle method (e.g., Kim, 2010; Sadarjoen, 1999). The nearly
closed contours are clustered with the same sign of vorticity and con-
ected as a time series when the center of the cluster falls within a spatial
range that is less than the typical translation distance. The identified eddy
time series with a persistency of less than 12 h is excluded from the analysis
(see Kim, 2010, for more detailed procedures and estimates of the stream
function and vorticity directly from the multiple radial velocity maps).

The individual PDFs of the diameters and Rossby numbers of the identified eddies, as well as their joint PDF, reveal that the dominant sizes of the eddies range from 5 to 10 km under the Rossby numbers of $-1$ to 2.5
(except for $|R_o| \leq 0.3$) (Figures 14a–14c). The vorticities of the identified eddies show asymmetry at high vortici-
ity values ($|R_o| > 1$) (Figure 14c), which can be explained by the fact that clockwise (anticyclonic) eddies with
high vorticity values are unstable, whereas counter-clockwise (cyclo-
nic) eddies with similar magnitudes of vortic-
ity are stable (e.g., Buckingham et al., 2016; Hoskins & Bretherton, 1972). In addition, the 10 day averaged means
and standard deviations of the number of eddies, normalized vorticity, and diameter do not show clear season-
ality (Figures 14d–14f). The significant variability of surface vorticity appears from early April to early May and
from mid-August to mid-November (Figures 11a and 11b), which correspond to the time periods that the sub-
mesoscale eddies with high vorticity are observed (Figure 14c). Considering the variability of the regional wind

![Figure 14.](image-url)
(Figure 4), the time periods that down-front winds are set up and their speeds are above a threshold (e.g., 2 m s\(^{-1}\) at the W1 station and 5 m s\(^{-1}\) at the W2 buoy) agree with the enhanced vorticity activities to some degree, which may imply that the interactions of the regional front and wind-driven circulations can lead the enhanced submesoscale processes (e.g., Thomas and Ferrari, 2008).

### 3.2.5. Variability of Mesoscale Circulation

The mesoscale variability in the study domain is represented by the time series of the reconstructed temperature profiles (Figures 5d, 5f, 5h, and 5j) and the SSHAs and SST along designated two sampling lines normal (n, PN–PS line) and parallel (p, PW–PE line) to the regional SPF (Figure 15).

The regional mixed layer depths, represented by density anomalies \(\Delta \rho = \rho - \rho_0\) of 25.5 and 26.5 kg m\(^{-3}\), show the variability at the seasonal frequency (SA\(_n\)) and its superharmonic frequencies (e.g., SA\(_3\), SA\(_5\), and SA\(_7\)) (section 2.3) (Figures 5d, 5f, 5h, and 5j).

The first-order finite spatial differences of the SSHAs and SST along evenly spaced axes allow us to identify where and when the mesoscale eddies and fronts occur (Figures 15c, 15d, 15g, and 15h), as they can be hidden in the dominant seasonal variability of the SSHAs and SST (e.g., Figures 15b and 15f). The spatial differences of the SSHAs can be an alternative to the geostrophic currents normal to the sampling line. The regional mesoscale boundary currents (e.g., NKCC and EKWC) are stronger in summer than they are in winter, but their seasonal differences are not very significant due to their semiannual variability (e.g., Kim & Min, 2008), which may support the enhanced variance in the energy spectra of the surface currents in summer (Figures 13b and 13d). The reduced wintertime energization of the mesoscale has been reported in the form of the slightly stronger variability of mesoscale eddies in summer than in winter (e.g., Qiu, 1999; Qiu & Chen, 2004; Sasaki et al., 2014), which can be explained by the unequilibrated process with intraannual variability based on observations in this paper (see Callies et al., 2016, for more details).

Conversely, the SPF is more active in fall than it is any other seasons, as identified by (1) the magnitude of geostrophic currents and the spatial gradients of SST and SSHAs to form the front and (2) the time duration to maintain the front (Figures 15c, 15d, 15g, and 15h). Based on the geostrophic currents normal to the chosen sampling axes (Figures 15c and 15g) and SSHAs (Figures 15a and 15e), the shear currents centered by the local maxima and minima of the SSHAs are identified (solid black arrows in Figure 15), and their temporal migrations along the sampling axes are marked (dashed black arrow in Figure 15). Although mesoscale eddies with lifetimes of less than 2 weeks are not marked explicitly in Figure 15, the mesoscale variability does not exhibit significant seasonal contrasts in summer and winter (Figure 15) (e.g., Kim & Min, 2008).

### 3.2.6. Interpretations of Energy Spectra

The potential drivers of submesoscale processes have been considered to be frontogenesis associated with mesoscale eddies (strain-induced frontogenesis), baroclinic instability in the mixed layer (mixed layer instability), turbulent thermal wind, and topographic wakes (e.g., Callies et al., 2015; McWilliams, 2016). Specifically, the mesoscale-driven surface frontogenesis energizes the submesoscale processes. When baroclinic instability within the mixed layer primarily drives the submesoscale flows, the decay slopes of the energy spectra in summer become steeper than those in winter, as the instabilities within the shallower mixed layer in summer are damped out easily. The topographically driven submesoscale processes may have nearly constant and persistent patterns of decay slopes of their energy spectra regardless of the season.

The regional coastal observations in this paper exhibit (1) mesoscale regional and boundary circulations and (2) submesoscale eddies with diameters of 5–10 km and effective Coriolis frequency ranging from \(\sim 1\) to 2.5. The weak seasonality in the energy spectra may be associated with the variability of the mixed layer depth, as modulated by seasonal superharmonic frequencies and regional mesoscale boundary currents (e.g., NKCC). Thus, these findings raise the possibility that submesoscale processes in this region could arise from a mixture of (1) regional low-frequency circulations or topographically linked shear currents and (2) baroclinic instability in the mixed layer with moderate seasonality. We are presently unable to isolate the dominant mechanism of the regional submesoscale processes among the above two mechanisms. However, baroclinic mixed layer instabilities can occur and grow in the presence of a mesoscale eddy field via their energization under mesoscale buoyancy gradients and strain fields (e.g., Callies et al., 2016; Mensa et al., 2013; Sasaki et al., 2014). Moreover, the dominant length scales of the regional submesoscale eddies appear in \(O(1–10)\) km. Thus, the regional submesoscale process can be driven by baroclinic mixed layer instabilities under the influence of mesoscale circulation and coastal boundaries.
4. Discussion

4.1. A Potential Theoretical Framework and an Injection Scale of Regional Submesoscale Turbulence

In the oceanic energy cascade, an injection scale is defined as the length scale either to delineate the inverse energy cascades and forward enstrophy cascades (e.g., Lee & Kim, 2018; Vallis, 2006) or to have a single zero-crossing wavenumber (an inverse of the separation length scale) in the kinetic energy fluxes (e.g., Frisch, 1995; Rhines, 1977; Scott & Wang, 2005; Soh & Kim, 2018) or both.

The spectral decay slopes of the wavenumber domain kinetic energy spectra of the surface currents become steeper from $k^{-5/3}$ to $k^{-3}$ or $k^{-5}$ at a scale of approximately 10 km (Figures 13e and 13f). However,
The kinetic energy flux $\Pi(k)$ can be estimated using the scale-by-scale energy budget equation in the one-dimensional wavenumber domain (e.g., Frisch, 1995; Rhines, 1977; Scott & Wang, 2005; Soh & Kim, 2018):

$$\Pi(k) = \langle u_x \cdot (u_x \cdot \nabla u_x) \rangle + \langle u_y \cdot (u_y \cdot \nabla u_y) \rangle, \quad (12)$$

where

$$u(x) = u_x(x) + u_y(x), \quad (13)$$

and $u_x$ and $u_y$ denote low-pass and high-pass filtered currents in the wavenumber domain, respectively. The kinetic energy fluxes are estimated with 6 hourly averaged surface current maps to maintain the primary signals of low-frequency, near-inertial, and tidal variances and to increase the signal-to-noise ratio. Only cases to have a single zero-crossing wavenumber are further analyzed to compute the bin-averaged kinetic energy fluxes and their PDF, which several cases in October and November of 2013 for both $S_x$ and $S_y$ show ambiguous tendency as opposed to the cases with steepened slopes (Figures 13e and 13f). Moreover, the wavenumber domain energy spectra of hourly geostationary ocean color imagery (GOCI)-derived chlorophyll concentration maps sampled in the same area at a spatial resolution of 0.5 km show the transition of the spectral decay slopes from $k^{-5/3}$ to $k^{-1}$ at a scale of approximately 6 km and from $k^{-1}$ to $k^{-3}$ at a scale of approximately 1 km (Figure 16a) (Table 1), which implies that the surface chlorophyll concentrations might be considered as a passive tracer under reasonable assumptions and careful analyses to exclude cases of nonphysical and biological interactions (see a companion paper Lee & Kim, 2018, for more details). Based on the energy spectra of the two kinds of surface submesoscale observations, the regional turbulent flows are more relevant to the QG turbulence theory rather than they are to the sQG and fsQG frameworks (e.g., Vallis, 2006; Tulloch & Smith, 2006).

Figure 16. (a) Wavenumber domain energy spectra of the hourly normalized GOCI-derived chlorophyll concentration maps sampled in the identical study domain are averaged for a period of 1 year (2013) $Q_x(k)$ for cross-shore direction (blue) and $Q_y(k)$ for alongshore direction (red). The wavenumber domain energy spectra of the chlorophyll concentration maps sampled in the cross-shore ($x$) and alongshore ($y$) directions are averaged in the alongshore and cross-shore directions, respectively. Gray axillary lines of the spectral decay slopes of $k^{-5/3}$, $k^{-1}$, and $k^{-3}$ are overlaid. The individual wavenumber ranges $[0.11 \leq k_x \leq 0.2 \text{ km}^{-1}]$ (gray), $[0.22 \leq k_y \leq 0.48 \text{ km}^{-1}]$ (orange), and $[0.6 \leq k_x \leq 0.95 \text{ km}^{-1}]$ (green) and dotted boxes are marked, and the relevant spectral decay slopes of $k^{-5/3}$, $k^{-1}$, and $k^{-3}$ are highlighted with corresponding thick lines within the dotted boxes, respectively. (b) A bin-averaged wavenumber domain kinetic energy flux of the surface currents for a period of 1 year (2013) when the estimated energy fluxes have a single zero-crossing and the probability density function of zero-crossing wavenumber. The surface currents are sampled within a black box as the subset of study domain in Figure 16a. (c) Eddy anisotropy ($\kappa$ in equation (15)) of the surface currents. As the eddy anisotropy becomes zero and one, the vector current is close to the isotropic and anisotropic conditions, respectively. The bottom bathymetry is contoured at 50, 100, 200, 300, and 400 m.
accounts for approximately 21% of total realizations. The rest of the estimated kinetic energy fluxes are classified into the cases having all positive and all negative values within the given wavenumber, which can be interpreted that the energy cascades occur below and above the range of the observational spatial scales (2–20 km in this paper) (see Soh & Kim, 2018, for more detailed interpretations of the kinetic energy fluxes estimated from observations). The bin-averaged kinetic energy flux of the surface currents and the PDF of the zero-crossing wavenumber show that the dominant injection scales range from 4 to 10 km (Figure 16b). These estimated injection scales overlap with the spatial length scales of the regional submesoscale eddies (Figures 14a and 14f) and are consistent with the baroclinic instability associated with the (seasonal) mixed layer (e.g., Brink, 2016; Callies et al., 2015; McWilliams, 2016).

In this region, the submesoscale processes may be associated with baroclinic instability in the mixed layer with moderate seasonality under the mesoscale circulation, including low-frequency circulation and topographically linked shear currents (section 3.2.6). Thus, the regional and boundary circulations need to be taken into account when modeling the submesoscale processes in this region, which was not included in Callies et al. (2016). The potential mechanisms of the regional submesoscale circulation can be investigated using regional numerical modeling with the combinations of boundary circulations and potential driving forces.

4.2. Anisotropy of Surface Circulation

The regional surface circulation is characterized by (1) the anisotropic variance ratio (Figure A1e) and the nonzero eddy anisotropy ($\kappa$ in equation (15); Figure 16c) of the surface current components, (2) the spatial structure of the variance in the primary frequency bands as a function of distance from the coast (Figure 7), and (3) the nonidentical kinetic energy spectra averaged in the cross-shore and alongshore directions (Figures 13c and 13d), which can be associated with the circulation constrained by coastal boundaries (e.g., coastline and bottom bathymetry).

The eddy anisotropy ($\kappa$) is defined as,

$$\kappa = \sqrt{\frac{(u'^2) - (v'^2))^2 + 4(u'v')^2}{(u'^2) + (v'^2)}.}$$

where $u'$ and $v'$ denote the time-varying current components with the temporal mean of $u$ and $v$ removed, respectively (e.g., Qiu et al., 2017; Stewart et al., 2015). As the eddy anisotropy reaches values of zero and one, the vector current approaches the isotropic and anisotropic conditions, respectively. The estimated eddy anisotropy of surface currents varies between 0.4 and 0.8, similar to the variance ratios of the surface current components (Figures A1e and 16c), which exhibit anisotropic surface circulation.

The clockwise near-inertial surface currents show onshore phase propagation and decreasing amplitudes (Figures 7c and 10). When the frictional effects due to bottom topography and shoreline become dominant, the onshore decreasing near-inertial currents can be consistent with coastal inhibition. The pure-circular clockwise near-inertial motions are the resonant ocean responses to broadband wind stress (or the minimum level of wind variance at the frequency of interest) and can be modified into elliptical motions due to sheared geostrophic currents or background vorticity (e.g., Whitt & Thomas, 2015). The regional superinertially shifted clockwise near-inertial surface currents are consistent with the near-inertial (internal) motions that propagate from their generation sites to the coast.

The anisotropy of the energy spectra can be related to (1) the anisotropic wind-driven responses constrained by coastal boundaries, (2) the near-inertial surface currents bounded by coastal boundaries, and (3) the coastally trapped waves and enhanced tidal currents near the coast, compared with the isotropic surface ocean responses in the unbounded open ocean.

4.3. Feasibility of the Use of HFR-Derived Surface Currents for Studies of Submesoscale Processes

The $O(1)$ m depth-averaged current observations in the upper ocean with a horizontal spatial resolution of $O(1)$ km and an hourly temporal resolution derived from coastal HFRs provide surface currents as well as their kinematic and dynamic quantities (e.g., stream function, velocity potential, divergence, vorticity, and strain rates). They can be used as resources in investigations of the geometric and physical properties of submesoscale eddies and fronts. This integrated analysis, which uses maps of concurrently observed SST, sea surface salinity, and chlorophyll concentrations at spatial and temporal resolutions with similar orders of
magnitude as those of surface currents, can provide insights that can be used in purely observation-based studies of submesoscale processes (Figure 1) (e.g., Lee & Kim, 2018).

Submesoscale observations of surface currents can provide a finely detailed picture of the variability of coastal circulation, particularly low-frequency coastal currents within 50–100 km of the coast (e.g., coastally trapped waves, ageostrophic coastal currents, and regional counter-currents near the coast) (e.g., Kim, 2010; Kim et al., 2013), diurnal circulation associated with the diurnal marine boundary layer (e.g., Hunter et al., 2007; Kim & Crawford, 2014), and near-inertial currents, including decorrelated spatial structures, phase propagations, and elliptical motions (e.g., Kim et al., 2014, 2015; Kim & Kosro, 2013). The stream functions of the observed surface currents can provide a quantity proportional to the SSHAs near the coast and quantify the coastal geostrophic currents, which can be a substitute of band-pass filtering to delineate the geostrophic and ageostrophic currents (e.g., Kim, 2010).

The strength of a radar signal diminishes as a function of distance from the shoreline, and intermittent missing data in the radial velocity map may introduce spatial biases and nonphysical properties into observations. However, the spatial structure (e.g., coherence and correlation) of the obtained surface currents in all available frequency bands or specific frequency bands and their degrees of statistical and dynamical consistency with the independent observations of other variables (e.g., wind, subsurface currents, and sea surface heights) can be good indicators with which to identify spurious responses to primary driving forces (e.g., Kim, 2010; Kim et al., 2009, 2015).

5. Conclusions

Here we examined the spectral contents of the observed submesoscale coastal surface currents with an hourly temporal resolution and a 1 km spatial resolution in the frequency and wavenumber domains, which were obtained from phased-array HFRs off the coastal region of Imwon, Republic of Korea, over a period of 1 year (2013). The primary variance of the observed surface currents appears in two frequency bands, including the low-frequency (longer than 2 days) and clockwise near-inertial frequency bands. The low-frequency surface currents are more consistent with the regional altimeter-derived geostrophic currents in summer than those in winter because the ageostrophic currents can be minimal associated with relatively weak wind conditions in summer. The wavenumber domain kinetic energy spectra of the surface currents become steeper, increasing from a slope of \( k^{-5/3} \) at a length scale of approximately 10 km to slopes between \( k^{-2} \) and \( k^{-3} \) at a length scale of 2 km with weak seasonality. These kinetic energy spectra exhibit (1) anisotropy depending on sampling directions of the surface currents, associated with the influence of anisotropic coastal boundaries (e.g., the shoreline and bathymetry), (2) weak seasonality resulting from persistent regional circulations and mixed layer depths modulated by seasonal superharmonic frequencies, and (3) an injection scale of \( O(10) \) km that is consistent with the dominant length scales of the regional submesoscale eddies and the zero-crossing wavenumber in the estimated kinetic energy fluxes. These findings suggest that the submesoscale processes in this region are primarily initiated by surface frontogenesis caused by regional mesoscale currents, including low-frequency circulations and topographically linked shear currents, then maintained by baroclinic instability in the mixed layer with moderate seasonality.

Appendix A: QAQC of the Radial Velocities

The sampling errors of the HFR-derived radial velocities can be quantified in terms of the standard deviations \( \lambda \) of the sum of paired radial velocities \( (r_a + r_b) \), which are obtained from closely spaced radial grid points (e.g., less than 10% of the range spacing) of independent HFRs, and their correlations \( \rho \),

\[
\lambda = \sqrt{(\langle r_a + r_b \rangle)^2},
\]

\[
\rho = \frac{\langle r_a r_b \rangle}{\sqrt{(\langle r_a^2 \rangle)\langle r_b^2 \rangle}},
\]

which can be considered as the uncertainty and SNR of surface current observations obtained using a local radar system, respectively (e.g., Lipa et al., 2006; Kim, 2015; Kim et al., 2008).
Kim et al. (2008), Chavanne et al. (2007), Kim (2015) assumed an isotropic current field ($a^2 = 1$, where $a^2 = \frac{\langle v^2 \rangle}{\langle u^2 \rangle}$ denotes the variance ratio of the vector components) and zero cross correlation between vector components ($b = 0$). However, we assume that the current components are anisotropic ($a^2 \neq 1$) and have nonzero cross correlation ($b \neq 0$). The time series of the vector current components can be formulated as follows:

$$v(t) = a^2 \beta u(t) + (1 - a^2) \nu(t),$$

and the variance of the noise time series $\nu(t)$ uncorrelated with the current components is determined as follows:

$$\langle \nu^2 \rangle = \frac{a^2 (1 - b^2)}{(1 - a^2)^2} \langle u^2 \rangle.$$

The standard deviations of the sum of the paired radial velocities and their correlations are presented as a function of the bearing angle difference and the three different range spacings [1 km (black triangle), 1.2 km (red box), and 1.5 km (blue circle)] at IMWS (Figures A1a to A1c). The effective spatial coverage area is marked with a gray contour in Figure A1c when both (1) the temporal data availability is more than 50% and (2) the estimated $\lambda$ and $\rho$ are close to the expected values. (d and e) Estimated $\beta$ and $a^2$. Note that Figures A1d and A1e share a colorbar, and the ranges of the individual variables are marked separately. (f–h) A joint PDF of $a^2$ and $\beta$ (log 10 scale) and their individual one-dimensional PDFs.
(Figures A1d and A1f). These conditions correspond to an SNR of 3.33, which indicates that the current variance is approximately triple the noise variance.

Within the study domain, the ranges of the variance ratios and cross correlations of the observed vector currents are shown in Figures A1f and A1h. The expected values of $\lambda$ and $\rho$ have a modified cosine shape, which is different from the exact cosine function used in previous studies (e.g., Chavanne et al., 2007; Kim, 2015; Kim et al., 2008) (Figures A1a and A1b). The estimated values of $\lambda$ and $\rho$ are plotted with darker and paler symbols (Figures A1a and A1b) when they follow and deviate from the expected values, respectively, and their sampling locations are marked in Figure A1c. The gray contour in Figure A1c denotes the area where the temporal data availability for the vector currents is greater than 50% and the estimated $\lambda$ and $\rho$ follow the expected values under varying $\sigma^2$. Based on the oppositely facing radial velocity observations, wherein the difference in the bearing angles is equal to 180°, the uncertainty of the surface current measurements using HFRs in this region is approximately 15 cm s$^{-1}$ (Figure A1a). Considering the uncertainty and SNR in other studies (e.g., Kim, 2015; Kim et al., 2008), the estimated uncertainty and SNR are of a similar order of magnitude.

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