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Lagrangian and Eulerian characterization of two counterrotating submesoscale eddies in a western boundary current

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Key points

1. Evidence of submesoscale ageostrophic filaments and eddies in a western boundary current from HF radar, moored, and satellite observations
2. Influence of the short-term wind variability on the surface currents, horizontal flow shear, frontal jet destabilization, and cyclonic eddy formation
3. Contribution of coastal eddies and submesoscale fronts to the cross-shelf dispersion, advection, and mixing of productive shelf waters

ABSTRACT

In recent decades, high-resolution ocean radar and satellite imagery measurements have revealed a complex tangle of submesoscale filaments and eddies, in the surface velocity, temperature and chlorophyll-a fields. We use a suite of high resolution data to characterize two counter-rotating, short-lived eddies formed at the front between the warm East Australian Current (EAC) and temperate coastal waters (30°S, Eastern Australia). In this region, submesoscale filaments and short-lived eddies are dynamically generated and decay at time scales of hours to days. Dominant cyclonic filaments of O(1) Rossby number formed along frontal jets and eddy boundaries, generating localized ageostrophic circulations at the submesoscale. Measurements of over-ocean wind direction and surface currents from high-frequency radars reveal the influence of the short-term, small-scale wind forcing on the surface circulation, enhancement of the horizontal shear, frontal jet destabilization and the generation and decay of the cyclonic eddy. By contrast, the anticyclonic eddy formation was most likely associated with EAC mesoscale instability and anticyclonic vorticity. Lagrangian tracks show that surface particles can be temporarily trapped in the eddies and frontal convergent zones, limiting their transport. Mixing between EAC-derived and coastal waters was increased along the frontal regions, and particles starting at the divergent regions around the eddies experienced significant dispersion at submesoscales. The cyclonic cold-core eddy entrained high chlorophyll-a shelf waters on its convergent side, suggesting spiral eddy cyclogenesis.

Keywords: ocean high-frequency radars (WERA), wind forcing, submesoscale eddies, East Australian Current, particle tracking, relative diffusivity

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1. Introduction

In recent decades, high-resolution current measurements and remote sensing imagery have begun to unveil the surface dynamics of the ocean at kilometer scales. Sea-surface color images [Pasquet et al., 2012; Alpers et al., 2013] and observations of surface currents by ocean radars [Roughan et al. 2005; Parks et al., 2009; Haza et al., 2010] and Lagrangian drifters [Poulin et al., 2013; Mantovanelli et al., 2012] reveal a complex tangle of eddies, fronts and filaments on scales of 1-100 km [Ferrari, 2011]. These coherent flow structures have a significant impact on coastal patterns of transport and mixing, where they regulate biogeochemical fluxes and the dispersal of pollutants and organisms [Lekien et al., 2005; Baird et al., 2011; Roughan et al., 2011].

Of particular interest are the dynamics of coastal eddies that form along western boundary currents because of their influence on biological productivity and fisheries [Everett et al., 2011, 2015; Mullaney and Suthers, 2013]. Persistent eddies can isolate water masses for prolonged periods, aggregate or juxtapose populations of organisms, serve as reproduction refuges, and affect biological connectivity [Owen, 1981; Largier, 2003; Roughan et al., 2011]. In addition, the cross-shelf transport of nutrients by coastal eddies can limit biological production in highly productive eastern boundary upwelling systems [Gruber et al., 2011] or enhance productivity in typically oligotrophic western boundary currents [Everett et al., 2015]. Less is known about the influence of short-lived submesoscale eddies on coastal productivity.

Here we analyze the evolution of two short-lived, submesoscale eddies in the East Australian Current (EAC) separation region, off Coffs Harbour (Eastern Australia, Fig. 1a), and investigate the response of these eddies to spatiotemporal variability of the wind forcing. In this region, the continental shelf is narrow and the mesoscale circulation is dominated by the EAC, which flows poleward as a swift jet of warm waters along the slope with variable current speeds and volume transport [Mata et al., 2000]. Intrusions of the EAC near the coast generate thermal gradients and intensify the baroclinic flow [Schaeffer et al., 2013, 2014a]. The local wind forcing has two main effects on the circulation in the region: south-westward blowing winds generate an offshore Ekman transport, coastal upwelling and a poleward jet on the shelf; conversely, north-eastward blowing
winds transport water onshore and favour downwelling and equatorward flow [Middleton et al., 1997; Schaeffer et al., 2013, 2014a; Rossi et al., 2014].

In this study, we combine a suite of remote and in situ measurements (high-frequency radar surface currents and wind direction, temperature and current profiles and satellite imagery) to generate hypotheses about the formation and decay of two counter-rotating structures, one cold-core cyclonic eddy and one warm-core anticyclonic eddy. Radar-derived maps of over-ocean wind direction [Wyatt, 2012] show variability over short temporal and spatial scales and suggest the influence of the wind on the generation of horizontal surface shear flow and frontal destabilization. Both Eulerian (vorticity, divergence, Rossby number) and Lagrangian (radar-based tracking, relative diffusivity of particles) approaches are applied to characterize the complex flow patterns and particle dispersion produced by the eddies.

In a related article, Schaeffer et al. [subm.] present results from an automated eddy detection algorithm applied to radar surface currents from the same region over a one year period and show that cyclonic eddies are generated on average every 7 days, and anticyclonic eddies occur less frequently. The focus of Schaeffer et al. [subm.] was on frontal eddies associated with meanders of the EAC that propagate at the inshore edge of the western boundary current over days or weeks independently of the wind stress. Most of the eddies were short-lived (life span < 2 days within the radar coverage) and a more detailed investigation on the role of the short-term wind-variability on their generation is required. This article investigates in detail the formation of two shorter-lived eddies from the same period, describing their evolution and response to wind variability using a variety of observational and analytical techniques, in order to highlight the hypothesis-generating power provided by combinations of data-types.

The article proceeds as follows. In Section 2 we describe the data and methodology used in this study. In Section 3, we describe the influence of the short-term wind variability (hours to days) on the circulation of this region, before discussing in detail the evolution of two submesoscale eddies and their response to wind variability. In Section 4, we discuss the Lagrangian properties and biological response of the eddies and hypothesize possible mechanisms responsible for their formation and decay. Section 5 summarizes our conclusions.

2. Data and Methods
2.1. HF radar measurements
A pair of land-based WERA phased-array high-frequency (HF) radars remotely measures surface currents in the top 0.9 m offshore Coffs Harbour (Eastern Australia, 30-31°S; Fig. 1a). This HF system operates at 13.92 MHz frequency with radial and azimuthal resolutions of 1.5 km and 10.4°, respectively [Wyatt et al., 2017]. Radial components (FV01 netcdf radial files; available at http://imos.aodn.org.au/imos) with intersection angles between 30 and 150° (good GDOP range), Bragg signal to noise larger than 10 and temporal coverage above 50% (over Sept. 2012 to Sept. 2013; Fig. 1a) were averaged over a 30 minute moving box and then combined to produce maps of the surface current field (u and v-components) on a rectangular grid mesh with a spatial resolution of ~1.5 km and a temporal resolution of 10 minutes [Wyatt et al., 2017]. All times for radar measurements are in UTC.

At each grid point, absolute current velocity components larger than their annual averages plus 5 times their standard deviations or larger than 2.5 m s⁻¹ were removed; the use of a Hampel filter (over 3 samples; 5 standard deviations) further improved the outlier removal. The u- and v-components were posteriorly smoothed over a 6-hour window, using a polynomial (2nd degree) least square fitting Savitzky-Golay algorithm [Savitzky and Golay, 1964], and short temporal gaps (less than 1 h) were linearly interpolated. The Savitzky-Golay filter replaces each measurement with the constant term in the polynomial that is found by weighted (using the data variances) least squares fitting to the data points within the averaged time span. The variance of the constant term, and hence that of the smoothed current, is determined from the original variances through the least squares fitting procedure [Wyatt et al., 2017]. Fig. 1a shows the HF radar standard deviations after the 6 h smoothing for the v-component of the flow, calculated as the square root of the averaged variances over Sept. 2012; the errors for the u-component were smaller and were not shown (more details on the HF radar error estimation are found in Wyatt et al., 2017). The data smoothing further reduced data noise while retaining the short-time response of the flow to high-frequency winds on timescales of 7-27 hours (Fig. 2a).

Tidal currents represented only 2% (in average) of the flow variability over a one-year period (Sept. 2012 to Sept. 2013) and could not be accurately separated from the wind-driven currents, which have strong diurnal and semi-diurnal signals and inertial oscillations in the Coffs Harbour region. As a consequence, the data were not de-tided. Finally, surface current components were spatially averaged using an overlapping window to produce a 3 km spatial resolution.
2.1.1. Over-ocean wind direction from HF Radar

Following the method of Wyatt [2012], the direction of the wind over the ocean was extracted from the two first-order Bragg peaks in the Doppler spectrum of the HF radar signal. These signals are backscattered from ocean waves with half the radar wavelength, one of which is propagating towards and the other away from the radar. These are short ocean waves that, except at very low wind speeds, are driven by the local wind. Wind directions are obtained from the relative amplitude of the two Bragg peaks from each of the two radars, applying a maximum likelihood fit of a hyperbolic secant function to the data at the cell of interest and the eight surrounding cells [Wyatt et al., 1997; Wyatt, 2012]. This process is repeated for all cells with sufficient first order signal to noise (>10 dB), providing maps of hourly averaged wind direction (but not wind speed) with a 3 km spatial resolution over the radar domain. This procedure is implemented in a software package provided by Seaview Sensing Ltd. Wyatt [2012] reports RMS differences between HF radar-derived and measured wind directions of 30-50°; however these measurements were not co-located.

Unit vectors were used to represent the HF radar wind field in the Figures (i.e. Fig. 3, Fig. 4 and Fig. 8) because the method can only estimate wind direction, not wind speeds. The snapshot of wind direction with best spatial coverage measured 1-3 h before the HF currents was taken to allow for the lagged response of the surface flow (as shown in Fig. 2c). A second data set was also used that provides both wind direction and intensity every 6 hours with a 12 km spatial resolution obtained from the ACCESS numerical weather prediction model made available by the Australian Bureau of Meteorology (http://www.bom.gov.au/nwp/doc/access/NWPData.shtml). ACCESS data closest in time to HF radar observation were used in each plot. The time lags between the two data sets ranged between 0.5-2.5 h. ACCESS speeds in the text are given as mean values over the domain plus or minus the standard deviations. Linear correlations between ACCESS velocity data (closest to Coffs Harbour station; about 4 km away) and the land-based wind measurements produced r² of 0.8 and 0.5 for the v- and u-components, respectively.

There are no other co-located measurements of over-ocean winds in the Coffs Harbour region taken at the same spatial and temporal resolutions of the HF radar observations. Comparison between wind direction measured by the radar (at the closest location to the Coffs Harbour weather station) and land-based directions gave a complex correlation coefficient of 0.52 and weighted average angular separation of 3.7. However, these two measurements were taken about 40 km away. Better agreement was found for comparisons between HF radar and ACCESS directions (Sept. 2012-
Sept. 2013), with complex correlation coefficients ranging from 0.63 to 0.81 in 70 locations within the domain. Averaged angular separations were less than ±20° for 91% of the comparisons, and had a maximum absolute value of 38°. Overall the two data sets consistently represent the shifts between northward and southward wind directions, which are of interest for this work.

2.2. Over-land wind measurements

Over-land wind data measured at Coffs Harbour (station number 59040 located at 30.31° S and 153.12° E) every 30 minutes were obtained from Australian Bureau of Meteorology. Wind speed and are reported using the oceanographic convention (direction the wind blows to).

2.3. In situ ADCP current and temperature profiles

Current data profiles were measured every 5 minutes by two bottom-mounted ADCP (RDI Instruments, 307.2 kHz; bin size of 4 m) moored on the shelf (70 m isobath, CH70) and (100 m isobath, CH100; Fig. 1a); thermistors installed on the moorings recorded water temperature every 8 m along the water column for depths below 10 m [more details in Schaeffer et al. 2013, 2014]. The ADCP data were corrected for magnetic declination, quality-controlled to remove records contaminated by side-lobes and with low signal to noise ratio, and placed into uniform depth strata [Mantovanelli et al., 2015]. Comparisons between the 6 h smoothed ADCP (at 10 m depth) and HF radar surface current speeds (Sept. 2012 to Sept. 2013) showed a good agreement for both CH70 and CH100 mooring locations (r²~0.6-0.9; p~0; RMS~0.1 m s⁻¹); the linear regression and joint probability distribution for the CH100 mooring comparison is shown in Fig. 1b.

2.4. Satellite images

We used MODIS AQUA level 2 images (resolution of ~1 km) of sea-surface temperature (SST, °C) and chlorophyll-a (mg m⁻³, from OC4 algorithm; http://oceancolor.gsfc.nasa.gov) sampled daily or twice a day, projected to Mercator projection, and taken usually less than 10 minutes apart from the HF radar current measurements. The SST images served as a proxy for surface water density changes, which are mainly driven by the thermal gradients [Schaeffer et al., 2014]. A total of 46 SST and 17 color images, available during September 2012, were used to identify daily variations on the extension of the EAC encroachment upon the shelf and of chlorophyll-a spatial distribution, respectively.
2.5. **Eulerian diagnostics**

Eulerian flow properties were calculated using centered differencing of the gridded surface current components \((u, v)\) in the east-west \((x)\) or north-south \((y)\) directions, using the 3 km resolution and the 6 h smoothed HF radar data. The two-dimensional surface divergence, \(\gamma = \frac{\partial u}{\partial x} + \frac{\partial v}{\partial y}\), implies vertical upwelling for divergent flows \((\gamma > 0)\) or downwelling for convergent flows \((\gamma < 0)\) by continuity. The local relative vorticity, \(\zeta = \frac{\partial v}{\partial x} - \frac{\partial u}{\partial y}\), has anticlockwise \((\zeta > 0, \text{ anticyclonic})\) and clockwise \((\zeta < 0, \text{ cyclonic})\) rotations in the southern hemisphere. Both properties \((\gamma, \zeta)\) were divided by \(|f|\) (Coriolis parameter) to facilitate comparison. The squared strain rate is given by

\[
S^2 = \left(\frac{\partial u}{\partial x} - \frac{\partial u}{\partial y}\right)^2 + \left(\frac{\partial v}{\partial x} + \frac{\partial u}{\partial y}\right)^2.
\]

An uncertainty of the HF radar currents of 0.02 m s\(^{-1}\) (maximum error in Fig. 1a) over three grid points (6 km) suggests that \(\gamma/|f| < \pm 0.045\) are indistinguishable from zero and were left blank in plots. The divergence of surface current velocities is a measure of vertical flux to/from the thin surface layer (top meter of the water column) sampled by the HF radar, which may result from upwelling, downwelling or vertical mixing [Kaplan and Lagier, 2006].

The ratio \(Ro = \zeta/|f|\) gives an estimate of the local Rossby number, keeping positive or negative values for anticyclonic or cyclonic rotations, respectively. The loss of geostrophic balance and generation of ageostrophic motions occur at large local Rossby numbers \((Ro \sim O(1))\), where the relative vorticity of the flow equals or exceeds the planetary vorticity, contrasting with typical \(Ro \ll 1\) observed for mesoscale flows [Mahadevan and Tandon, 2006; Thomas et al., 2008]. Localized flow regions with \(Ro\) of \(O(1)\) or larger are dynamically defined as submesoscale flows [Thomas et al., 2008]. Submesoscale filaments typically form along fronts and at the edges of eddies through nonlinear instabilities of the mesoscale currents [Gula et al., 2015]. Narrow lines of \(Ro \geq 1\) are referred here as submesoscale filaments with strong ageostrophic circulation, and classified as cyclonic \((Ro \leq -1)\) and anticyclonic \((Ro \geq 1)\).

2.6. **Lagrangian analysis**

The 2D HF radar currents were inputted to a particle tracking model to compute particle positions over time; the model uses bilinear interpolation of the velocity field on the 3 km resolution grid at each time step (10 minutes), an Euler Predictor-Corrector numerical method and projection of
the tracked positions on the earth ellipsoid [Mantovanelli and Heron, 2012b]; this model had an accuracy of ~4 km after two days of tracking when tested against surface drifters [Mantovanelli et al., 2011]. Particles were released on a square grid with a grid spacing of 300 m in the HF radar domain and followed for 18 h. The relative dispersion ($D^2$) was calculated from the squared separation of the particle pairs, $D^2_{ab}(t) = (x_a(t) - x_b(t))^2 + (y_a(t) - y_b(t))^2$, where the indexes ($a$, $b$) refer to each distinct particle in the pair. The relative diffusivity was calculated for each pair of trajectories as:

$$K_{R_{a,b}} = \frac{1}{4} \frac{d}{dt} D^2_{a,b}(t)$$ [Klocker et al., 2012].

3. Results

3.1. Wind variability and influence on the surface circulation

We analyzed a one-year time series of wind speed recorded at the Coffs Harbour weather station (Sept. 2012-Sept. 2013) to characterize the temporal variability of the wind. Multitaper rotary spectra [Lilly, 2016] detected two main peaks at the semi-diurnal and diurnal signals but more pronounced in the $u$-component of the wind velocity (Fig. 2a). Analysis of the HF radar current spectra at Coffs Harbour mooring locations also showed diurnal and semi-diurnal peaks both for surface (HF radar) and subsurface (ADCP) measurements [Wyatt et al., 2017]. Wind varied in direction almost daily over the analyzed year, with a maximum persistence in each direction of less than 6 days (not shown).

Surface currents measured by HF radar represent a sum of geostrophic currents (mesoscale circulation), wind-driven ageostrophic currents (such as Ekman drift), tides and other contributions [Rio and Hernandez, 2003; Tokeshi et al., 2007]. Wind-driven currents near the surface are typically 1–4% of the wind speed at an angle of 0-40° to the wind direction [Ardhuin et al., 2009; Chang et al., 2012]. We evaluated the short-term (hours to a couple of days) wind influence on the surface circulation using the lagged cross-correlation between the wind and current velocity components [following Emery and Thomson, 2004]. In Sept. 2012, maximum correlation coefficients ($\rho_{xz}$) between north-south winds given by the ACCESS model and meridional HF radar surface currents of 0.6-0.7 were observed nearshore and in the slope region but the correlation deceased to a minimum of 0.2 at the farthest offshore site (Fig. 2b); these results clearly show the higher influence of the wind forcing on the surface circulation near the coast during this period.

We also calculated the lagged cross-correlations between the land-based winds (measured at the Coffs Harbour station) and both surface (HF radar data; top 0.9 m) and subsurface (ADCP data;
10 m depth) currents extracted at the CH70 and CH100 mooring locations (indicated in Fig. 2b); results were similar at both locations and also similar whether or not land-based measurements or ACCESS wind data were used. Correlations between north-south winds and meridional currents are shown in Fig. 2c and between east-west winds and zonal currents in Fig. 2d. Based on the 47 sequential SST images, we split the data analyses in two periods: (i) Sept. 1-19, when the main EAC branch was offshore the radar domain (solid lines in Fig. 2c,d; EAC offshore) and (ii) Sept. 20-30, when the main EAC branch encroached upon the shore (dotted lines in Fig. 2c,d; EAC nearshore).

Correlation coefficients ($\rho_{xy}$) between meridional winds and north-south surface HF currents were ~0.75 under weak EAC influence (Sept. 1-19) and reduced to ~0.55 as the EAC influence increases (Sept. 20-30). Surface currents presented a time lag response to the wind forcing of 1-2 h. Similarly, meridional correlations for the ADCP currents dropped from ~0.66 (Sept. 1-19) to ~0.32 (Sept. 20-30) but subsurface currents had a slower response (~3-5 h; Fig. 2c). Correlations between the east-west wind and current components were weaker ($\rho_{xy} < 0.4$) with a diurnal periodicity nevertheless they showed the same reduction pattern between the two periods (Fig. 2d). All correlations were higher near surface (HF radar currents) than subsurface (ADCP currents). A higher contribution of the wind-induced circulation near the surface is expected, since the wind-driven Ekman currents decay exponentially with depth [Graber et al., 1987].

These results clearly demonstrate that the importance of the wind forcing on the surface circulation increases at times and in regions less influenced by the mesoscale currents (EAC).

3.2. Relevance of coherent flow structures (eddies and submesoscale filaments)

Visual inspection of the HF radar surface currents (6 h smoothed, 3 km resolution), Rossby numbers and SST images over one-year period (Sept. 2012 – Sept. 2013) revealed that submesoscale filaments (3-30 km wide) often formed along the frontal jets and eddy boundaries associated with regions of high flow strain. The histogram of Rossby number for observations taken between Sept. 2012 and Sept. 2013 (25 million data points) shows that the distribution is notably skewed (skewness = -1.38), indicating the dominance of cyclonic filaments (areas with $Ro \leq -1$) which covered up to 30% of the radar domain (not shown).

Schaeffer et al. [subm.] carried out a statistical study of the same region using an automated eddy-tracking algorithm and identified 40 cyclonic eddies over the period between Sept. 2012 and Sept. 2013. At least six of these features appeared to be frontal eddies propagating along the EAC,
independently of the wind stress. Of the other structures, Schaeffer et al. found that the longest living eddies (up to 6 days under the radar coverage) were stalled by northward winds, while 34 other structures occurred during mean northward wind. Anticyclonic eddies occurred more rarely (16 in total over the analyzed year).

Here we address in detail the influence of wind forcing on two short-lived eddies, one cyclonic cold-core eddy (CCE) and one anticyclonic warm-core eddy (WCE) formed during Sept. 2012. These eddies lasted ~2 days from the time the flow started meandering, forming a distinct circular or elongated vortex, until they dissipated. The eddies were formed in the presence of frontal jets (with thermal contrast of 2-5°C) and at times of high wind variability that increased the horizontal flow shear as described in Section 3.2.1 (CCE) and Section 3.2.2 (WCE).

3.2.1. Cyclonic cold-core eddy (CCE)

We use SST images and HF radar maps of surface currents and wind direction between 10 to 15 Sept. 2012 to illustrate the intrusion of a warm jet in the Coffs Harbour region, its interaction with shifting winds and the development of a cyclonic cold-core eddy inshore of the jet (Fig. 3-5). The frontal jet consisted of a tongue of mixed waters (SST ~19-21°C) that detached from the offshore EAC main jet (inset Fig. 3a).

On Sept. 10, the jet was clearly visible as a 30 km wide warmer region (~19-21°C) of surface intensified southward flow (speeds up to 0.6 m s⁻¹ and kinetic energy up to 0.18 m² s⁻²) contrasting the colder shelf waters (~16-18°C; Figs. 3a,e and Fig. 6a). Filaments (~3-9 km wide) of cyclonic (anticyclonic) relative vorticities developed on the west (east) sides of the jet axis (Figs. 3b,e) associated with regions of high strain (S/|f| up to 0.8; Fig. 3e). The surface flow was strongly convergent (γ/|f| up to -0.8) in the north portion of the jet (30°05′ S-30°12′S; Fig. 3c,e), indicating downwelling and the onset of frontogenesis. Southward winds were favoring the jet (Fig. 1c) and HF radar/ACCESS wind maps show dominant, moderate (mean speed -5.6 ± 1.8 m s⁻¹) south-southwest winds within the domain (Fig. 3d). Thomas and Lee [2005] pointed out that frontogenesis is strengthened by down-front winds, that is, winds blowing over a baroclinic zone in the direction of the surface currents.

On Sept. 11, radar/ACCESS data show weak (mean speed < |2.2| ± 2.3 m s⁻¹) winds with spatially variable intensity and directions within the domain (not shown), and the coastal weather station recorded a shift from southward to weakly northward winds (Fig. 1c). At this stage, the frontal

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jet became unstable and meandered and surface currents flowed in opposing directions within different regions of the domain. On Sept. 12, winds rotated again towards south (Fig. 1c) and the jet re-aligned with the alongshore direction.

On the beginning of Sept. 13, the southward winds strengthened considerably presenting approximately uniform direction and intensity over the whole domain (mean speed of -10.0 ± 2.8 m s⁻¹ in Fig. 4a) and resulted in a mostly southward flow (Fig. 4d); the front weakened and Rossby numbers were overall reduced, except for a region of strong cyclonic vorticity in the north portion of the domain (Fig. 4d).

On the second half of Sept. 13, a complex circulation pattern (Fig. 4e, 14: 25 UTC) developed as the winds reversed towards the north again (Fig. 1c; mean speed of +7.0 ± 9.9 m s⁻¹ in Fig. 4b) and a cyclonic eddy formed in the northern portion of the domain (Fig. 4e). Strong northward winds persisted until the first half of Sept. 14 (mean speeds of +7.9 ± 1.9 m s⁻¹ in Fig. 4c,g), generating intense onshore currents and submesoscale ageostrophic circulation (Ro ~ -1) on the eastern flank of the eddy subjected to high strain (Fig. 4f). At this stage, colder coastal waters were entrained on the eddy core and the warm jet surrounded it (Fig. 5a). Surface convergence dominated on the north-eastern flank and divergence on the south-western flank of the eddy (Fig. 5b).

Profiles of water temperature and current speeds taken in the warm (CH70) and cold (CH100) sides of the frontal jet (south-west portion of the CCE) showed that: (i) the current reversed from southwest (before CCE) to northwest during the CCE formation (Fig. 6a) and (ii) the temperature of the mixed layer (top 40 m) dropped by ~2°C on the cold side of the front (CH100; Fig. 6b) and increased at the CH70 mooring as the warmer waters that surrounded the CCE reached it (not shown). The CCE decayed (Fig. 4k,l) as the northward wind speeds decreased on Sept. 14-15 (Fig. 1c; mean speed of +2.3 ± 1.3 m s⁻¹ in Fig. 4h,i).

Histograms of HF radar surface current direction are shown for periods (i) with northward winds opposing the frontal jet (Fig. 5c) and (ii) southward winds favoring the frontal jet (Fig. 5d) both during the frontal jet and CCE events (black bars in Fig. 5c,f) and the whole month of Sept. 2012 (blue bars in Fig. 5c,f). Westward currents (240-360°) were generated under northward winds (Fig. 5c) while the surface flow was mainly restricted to the south direction (160-200°) under southward winds (Fig. 5d). These figures clearly show a higher variability of the current direction, i.e. 1.5-2.0 times higher standard deviation and 1.4-1.6 smaller kurtosis, under northward winds (opposing the frontal jet; Fig. 5e) than under southward winds (favoring the frontal jet; Fig. 5d).
3.2.2. Anticyclonic warm-core eddy (WCE)

On 20-23 Sept. 2012, the EAC approached the shore and its detachment region progressively migrated south (Fig. 7a-c). A WCE formed on Sept. 20-21 (Fig. 8d-f,j-k). During this period, weak northward winds (mean speed of +2.2 ± 1.7 m s\(^{-1}\) in Fig. 8a) shifted to southward with weak (mean speed of -2.5 ± 1.1 m s\(^{-1}\) in Fig. 8b) to moderate speeds (mean speed of -6.5 ± 3.0 m s\(^{-1}\) in Fig. 8c,g) and back to moderate northward winds (mean speed of +6.0 ± 1.7 m s\(^{-1}\) in Fig. 8h,i).

On day 21, the WCE was fully developed (diameter ~60 km; Fig. 7b,e and Fig. 8j) on the EAC branch that partially intruded on the radar area. An elongated tongue of cold water was visible between the EAC main jet and its branch (inset Fig. 7b). A strong horizontal shear was observed between the southward flow (kinetic energy up to 0.25 m\(^2\) s\(^{-2}\)) of cyclonic vorticity on the shelf (Ro up to -0.6), and the eastward turning EAC branch with anticyclonic vorticity (Ro up to 0.6; Fig. 7g and Fig. 8j). Regions of surface divergence and convergence alternated within the domain (Fig. 7e). A thermal contrast of ~5°C existed between the cold shelf waters and warm EAC branch (Fig. 7b,g), with high strain (\(S/f\) \(\approx\) 0.7) on the cold side of the front on the slope (Fig. 7g). Both current profilers sampled within the frontal jet to show strong surface-intensified southward flows (e.g. CH100; Fig. 6a) associated with warmer waters on the slope (CH100 mooring; Fig. 6b) and ~2-3°C colder waters on the shelf (CH70 mooring; not shown).

On the second half of Sept. 21, the WCE migrated a few kilometers southwest and became elongated between the strong southward flow on the slope and the northward flow that intensified in open waters (Fig. 8k) as winds rotated again to the north-northeast (mean speed of +6.0 ± 1.9 m s\(^{-1}\) in Fig. 8h). The eddy finally decayed on Sept. 22 (Fig. 8l) when a dominant onshore flow developed under dominant northward winds (Fig. 8i). On Sept. 23, SST images show the encirclement of the cold tongue by the warmer waters (inset Fig. 7c), as the EAC moved further south and encroached on the shelf under south-westward winds (Fig. 1c).

4. Discussion

4.1. Influence on the Lagrangian transport and particle dispersion

We demonstrated that coherent flow structures (such as submesoscale fronts and eddies) are formed and dissipate in the region offshore Coffs Harbour at time scales of hours to days in response to the combined influence of the EAC and wind forcing. Spatial and temporal variations of the wind
direction and intensity and of the extent of the EAC encroachment upon the coast generate complex circulation patterns and the interaction between the frontal jet and winds can potentially aid the generation and dissipation of coastal eddies and submesoscale fronts. We applied the semi-empirical Lagrangian tracking to evaluate the short-term influence of these coherent flow structures on surface transport and particle dispersion.

Lagrangian particle tracking showed that particles accumulated in zones of strong flow convergence formed along frontal regions (Fig. 9a and Fig. 3c) as well as inside both the CCE (Fig. 9b and Fig. 5b) and WCE (Fig. 9c and Fig. 7e), and ~26-36% of the particles had left the domain after 18 h of simulation. Particles travelled south and offshore with the frontal jet (Fig. 9a). For the CCE particle tracking, a large number of particles were advected from the coast and concentrated in convergent regions inside and surrounding the eddy, with the remaining particles transported onshore or southward over the slope (Fig. 9b). Particle advection by the WCE shows that particles released on the northwest portion of the domain were either advected southward by the strong frontal jet flow or offshore (Fig. 9c).

The relative diffusivity ($K_R$) of a given particle pair depends on the particles’ starting locations, pair orientation and time. Here, pairs were formed between particles initially located at neighboring grid points in both the east-west and north-south directions. A sense of how the submesoscale eddies influenced the spatial pattern of particle dispersion can be obtained by plotting relative diffusivities $K_R$ (after 18 hours) at the starting location of the pair (Fig. 9d,e,f). Positive and negative values of $K_R$ (averaged over initial pair orientations) indicate particle divergence and convergence, respectively. Relative diffusivities were low ($|K_R| < 1 \text{ m}^2\text{s}^{-1}$; Fig. 9g,h,i) within most of the domain, with alternating divergent ($K_R > 0$) and convergent ($K_R < 0$) regions.

The positive relative diffusivities observed for the three events analyzed here (mostly below 5 m$^2$ s$^{-1}$; Fig. 9g,h,i) agree with the values estimated by Nencioli et al. [2013] for submesoscale fronts; these authors found that 70% of the eddy diffusivity values ranged between 0.4 and 5 m$^2$ s$^{-1}$. Maximum positive $K_R$ values of 12 m$^2$ s$^{-1}$, 53 m$^2$ s$^{-1}$ and 137 m$^2$ s$^{-1}$ were estimated for the frontal jet, CCE and WCE events, respectively. The WCE was more dispersive than the other two structures, with more instances of $K_R$ above 1 m$^2$ s$^{-1}$ (39%) than both the frontal jet and CCE (21-22%). This is likely due to the larger size of the anticyclonic eddy and the high kinetic energy of the frontal jet on its western boundary.

Narrow ridges of high positive diffusivity (i.e. $K_R > 1 \text{ m}^2\text{s}^{-1}$; red tones in Fig. 9d,e,f)
occurred along both east and west frontal jet sides (Fig. 9d), the south and east CCE flanks (Fig. 9e) and the western WCE boundary (Fig. 9f), which corresponded to starting locations of particles subjected to more intense spreading and dispersion. For instance, particle pairs released on the western flank of the WCE had larger relative diffusivity due to the separation of particles recirculating within the eddy, compared with those travelling consistently southward along the slope. By contrast, divergence on the southern flanks of the CCE increased the dispersion, mostly in the north-south direction but also in the across-shelf direction. This highlights the importance of the submesoscale frontal regions and eddies for enhanced mixing and across-shelf dispersion.

Radar-derived convergent flows indicate locations where the surface mixed layer is thickening and the surface waters are being subducted [Kaplan et al., 2006]. Conversely, surface divergence indicates regions of flow separation where surface waters are replaced by water that upwells from below. In these Lagrangian simulations, virtual particles are advected by the surface currents only and cannot move in the vertical direction. However, the existence of convergent or divergent zones has important implications for real particle dispersion as surface particles accumulated in convergent fronts (red tones in Fig. 3c, Fig. 5b and Fig. 7d,e,f) could be potentially mixed downwards. Conversely, upwelled water from divergent regions will be subjected to enhanced dispersion.

4.2. Biological response

These three coherent flow structures (frontal jet, CCE and WCE) exerted an influence on the surface chlorophyll-α distribution, as highlighted by 17 sequential ocean color images taken in Sept. 2017. The frontal jet advected productive coastal waters southward from the EAC separation zone (~29°S) up to the Coffs Harbour region (~31°S; Fig. 10a) over a period of three days; the cyclonic eddy entrained and transported shelf productive water offshore (Fig. 10b) and the chlorophyll-α concentrations were enhanced along the submesoscale fronts contrasting with the oligotrophic EAC waters (Fig. 10c).

On Sept. 12 (before the CCE formation), elevated chlorophyll-α concentrations were restricted to the near-shelf region (onshore the 50 m isobath). This enhanced productivity was likely associated with coastal upwelling driven by strong southward winds on Sept. 12 (Fig. 1c), a typical phenomenon in the region [Rossi et al., 2014]. At the time the CCE was formed, the productive shelf waters were entrained and advected ~40 km offshore by the frontal jet, elevating the chlorophyll-α
concentrations inside the CCE (Fig. 10b). This offshore advection is also seen in the Lagrangian simulations for particles starting offshore the 20 m isobath (Fig. 9b and Fig. 10b). A zone of intermediate chlorophyll-a is visible around the CCE boundary, likely resulting from the mixing of high chlorophyll-a waters with the oligotrophic waters surrounding the eddy (Fig. 10b). This is consistent with the overall increased dispersion (positive $K_R$) within the eddy and submesoscale fronts (Fig. 9e). Two days later, chlorophyll-a concentrations were still high in the region previously occupied by the CCE.

However, chlorophyll-a concentrations were much reduced (<0.3 mg m$^{-3}$) over the HF radar domain in the next available image on Sept. 20 (start of the WCE formation), as the oligotrophic EAC waters encroached on the shelf. Unfortunately, no ocean color imagery was available for the WCE, but it was unlikely to have elicited as strong a biological response as that of the CCE since the Sept. 23 satellite image (Fig. 10c) also shows oligotrophic waters near the coast except for relatively higher chlorophyll-a concentrations (~0.4-0.6 mg m$^{-3}$) associated with the submesoscale cold filaments (Fig. 7c). Submesoscale filaments can generate localized areas with strong ageostrophic circulations and vertical motions (O(100) m d$^{-1}$), playing an important role in local mixing and productivity [Lévy et al., 2012].

The interaction of eddies with a shelf-break current may account for a substantial portion of net cross-frontal exchange [Cenedese et al., 2013]. The two short-lived frontal eddies observed here could potentially facilitate the across-shelf exchange of marine organisms (such as larvae and eggs) between shelf and oceanic waters. The CCE promoted the offshore movement of nutrient-rich shelf-waters and the inshore movement of warmer EAC waters as well as temporarily trapped particles within the eddy and convergent frontal regions, limiting their southward advection. The WCE advected shelf waters offshore, increasing the across-shelf dispersion of a number of particles (and potentially marine organisms) initially located near the shelf, which are eventually reabsorbed by the EAC main jet and transported southward.

In a related paper, Schaeffer et al. [subm.] show that small-scale frontal eddies form regularly in this region. Recently, Roughan et al. [subm.] extensively sampled a similar cyclonic frontal eddy in the same region and found that the eddy was considerably more productive and had a higher proportion of coastal larval fish species than the surrounding waters. These studies and the results presented here suggest that short-lived submesoscale eddies may play an important role in productivity along the coast of south-eastern Australia.
4.3. Processes controlling submesoscale eddy formation and decay

Previous studies have shown the influence of wind for the generation of both: (i) highly non-linear cyclonic eddies ($|\text{Ro}| > 1$) linked to a wind burst event [Alpers et al., 2013] and (ii) anticyclonic eddies associated with baroclinic instabilities of growing flow meanders set up by shifts in the wind direction [Hansen et al., 2010] or bathymetric constraint of a wind-driven barotropic flow [Schaeffer et al., 2011].

The high-resolution radar-derived maps of wind direction used in this study permit a more detailed analysis of the role of spatial variation of wind direction on the formation, evolution, and decay of submesoscale eddies. We emphasize that, in the absence of a full, time-dependent, three-dimensional picture of the flow within the domain, such an analysis will necessarily be speculative. However, the dynamical scenarios discussed in this section are consistent with the radar-derived measurements of surface currents and wind direction. Moreover, the analysis demonstrates the potential of combinations of observational and analytical techniques to generate hypotheses about the driving processes in these ephemeral, highly dynamic flow features.

4.3.1. Cyclonic cold-core eddy (CCE)

A plausible mechanism for the generation of the short-lived cyclonic eddy described here was instability of the surface frontal jet accelerated by northward winds. Although the intruded frontal jet originally had a cyclonic relative vorticity (Fig. 4d; northern portion), the jet deflection towards the shore was boosted by the onshore flow generated under strong northward winds (Fig. 4f). This increased the horizontal velocity shear and, at the same time, stretched the meandering front, elevated the strain and resulted in localized submesoscale ageostrophic circulation on the eddy's eastern boundary (Fig. 4f,j). Cyclonic (anticyclonic) relative vorticities occurred on the cold inshore (warm, offshore) sides of the meandering frontal jet (Fig. 4f,j).

The CCE formation is consistent with theoretical models of spiral eddy cyclogenesis. The spiral eddy originates from an ageostrophic frontal preconditioning phase with strong cyclonic shear on one side of the front and weak anticyclonic shear on the other side, followed by shear (barotropic) instability that winds up the front [Munk et al., 2000] or baroclinic instability associated with frontogenesis on streaks of strong cyclonic shear and convergence [Eldevik and Dysthe, 2002].

Our hypothesis is that wind stress opposing the frontal jet can promote and/or accelerate
frontal destabilization at short timeframes (hours to days) and submesoscale spatial scales, aiding the generation of coastal cyclonic eddies. In the Coffs Harbour region, the frontal jet flows predominantly southward and the opposing northward winds can potentially destabilize the jet while southward winds strengthen it (as suggested in Fig. 3-5). This pattern is apparent in histograms of the surface current direction under northward and southward winds (Fig. 5c,f). The distribution of surface current direction under northward winds shows greater variability while the distribution of surface current direction under southward winds is largely confined to a narrow band. This suggests that interactions between the wind and frontal jet can increase the spatial variability of surface currents and generate complex circulation patterns.

Winds were mostly northward during the CCE formation with speeds of up to 8-11.1 m s\(^{-1}\) (Fig. 1c; orange line and maximum speed in Fig. 4b,c,g), this could potentially generate wind-driven currents of the order of 0.3-0.45 m s\(^{-1}\) (assumed 4% of the wind speed; Arduhin et al., 2009; Chang et al., 2012) deflected to the left (northward/onshore) of the wind direction. The northward currents opposed the southward jet (speeds ~0.3 to 0.6 m s\(^{-1}\); Fig. 6a on Sept. 13) and deflected the jet onshore. However the eddy started to decay as the wind intensity decreased (Fig. 4h,i) and the southward jet flow dominated again (Fig. 4k,l, end of Sept. 14 and Sept. 15). The entire scenario is more complex because both wind direction and the EAC influence vary spatially and temporally.

### 4.3.2. Anticyclonic warm-core eddy (WCE)

Several factors could have contributed to the formation and decay of the anticyclonic warm-core eddy, including the presence of submesoscale ageostrophic circulations, mesoscale baroclinic instabilities of the growing EAC jet meander and to some extent the high variability of the wind direction. A higher across-shelf thermal contrast (~5°C) was observed between the cold shelf and the warm EAC waters during the formation of the WCE (Fig. 7b). Relative vorticities of opposite sign occurred on either side of the energetic southward flowing frontal jet, with high cyclonic vorticity on the cold side of the frontal jet and anticyclonic vorticity on the warm side (Fig. 7g and Fig. 8j). Macdonald et al. [2013] showed that a mesoscale WCE started in a state of disequilibrium with a higher cyclonic vorticity ring outside at the density front between the EAC and the eddy and that the eddy grew by transfer of vorticity and mass from the EAC into the eddy.

Another possible generation mechanism for the WCE is associated with mesoscale instabilities of the EAC jet. A growing, wavelike distortion of the EAC jet was observed during the
formation of the WCE, leading to the intrusion of a colder tongue between two EAC warm branches (inset Fig.7b). A slantwise exchange of cold and warm waters is often generated across a heavily meandering jet, which eventually pinch off to form eddies when meanders amplify [Williams and Follows, 2011]. Frontal wave growth can result from different processes, such as baroclinic instability [Barth, 1989, 1994], combined local baroclinic/barotropic instabilities [Lozier et al., 2002] and interactions of an unstable jet with the coastline and bottom topography [Witter and Chelton, 1998; Lozier and Reed, 2005].

The wind stress during the WCE formation was predominantly southward (70% of the time) and weaker than that observed during the CCE (Fig. 1c, green line and mean speed between +2.2 to -7.2 in Fig. 8a,b,c,g). Therefore, it is unlikely that the frictional forcing induced by the wind alone could explain the current vorticity, and the EAC anticyclonic vorticity was most likely the primary driver for the WCE formation. The wind influence was more evident during the WCE decay (towards the end of Sept. 21), as the eddy was stretched between the southward flow on the shelf and the northward flow generated offshore (Fig. 8k) when the winds rotated towards north (mean speeds of 6.0 ± 2.0 m s⁻¹ in Fig. 8h). The eddy subsequently dissipated as moderate northward winds persisted (Fig. 8l), generating a mostly onshore flow (Fig. 8l).

5. Conclusions

We used a suite of high-resolution measurements to analyze the formation and decay of two counter-rotating submesoscale eddies (one cyclonic cold-core eddy and one anticyclonic warm-core eddy) in the East Australian Current separation zone and to generate hypotheses about the mechanisms controlling their dynamics. Spatial maps of over-ocean HF radar wind direction provided insight into the short-term, small-scale wind influence on the frontal jet and eddy dynamics. The analysis of the Eulerian flow properties and Lagrangian particle dynamics depicted the structure of the frontal eddies and show the influence of the eddies on surface transport and particle dispersion.

The near surface circulation changed over small spatial scales (a few kilometers) and short timescales (hours to days) due to the combined effect of the ageostrophic wind-driven flow and the mesoscale EAC current. The intensity and direction of the resultant surface flow was shown to depend on the relative strength and direction of these two flow components. The wind-driven currents are likely to be stronger either inshore or offshore of the frontal jet where they encounter less resistance from the opposing flow that increases the horizontal shear.
We hypothesize here that northward winds opposing the frontal jet aided the frontal jet meandering and the generation of the short-lived cyclonic eddy (Fig. 4 and Fig. 5). By contrast, the warm-core eddy (Fig. 7 and Fig. 8) formed under relatively weaker, southward winds and it is unlikely that the frictional forcing induced by the wind alone could explain the current vorticity; the EAC anticyclonic vorticity and its mesoscale instability most likely drove the warm-core eddy formation while winds aided the eddy dissipation. Our observations suggest that complex interactions between the frontal jet and winds near the coast are important for the understanding of the dynamics of submesoscale fronts and eddies in the EAC separation region.

Lagrangian particle tracking and relative diffusivities show that a large number of particles can be temporarily trapped at convergent fronts and within the core of eddies, limiting their transport to other areas. However, some particles starting in the divergent zones around the eddies experience strong dispersion at the submesoscale. Mixing between EAC-derived and coastal waters was increased along the submesoscale fronts, and satellite images suggest increased chlorophyll-α concentrations in these locations. The frontal jet advected productive shelf waters southward along the shelf and the CCE entrained biologically rich shelf waters, elevating chlorophyll-α concentrations on the slope and offshore regions, which were mixed with the oligotrophic EAC waters. Thus, short-lived coastal eddies and fronts may play an important role in the across-shelf transport of properties and marine organisms and mixing.

This paper demonstrates the power of and the need for the combination of high-resolution data from multiple platforms (such as HF radar, satellite images, ADCP) to analyze the dynamics of submesoscale eddies and fronts with short life span (hours to days) in western boundary currents systems. In addition to their power in generating and testing hypotheses about coastal ocean dynamics, data combinations such as those discussed here are useful for informing multi-platform sampling strategies; for indicating where the present technology, sensors and data products are lacking; and for providing impetus for the improvement of high-resolution models. In particular, the lack of high spatial and temporal resolution wind velocity data over the ocean prevents the separation of the wind-driven surface currents from the geostrophic flow and the evaluation of the role of the frictional forcing induced by the wind alone on the current vorticity. High-resolution data of wind direction from HF radar provided insights on the wind spatial and temporal variability and the interaction between submesoscale fronts and wind. The development of the techniques for the extraction of wind speeds from HF radar measurements would provide valuable data for the
understanding of the coastal dynamics at submesoscale.

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References


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Figure 1: (a) HF radar domain offshore Coffs Harbour (Eastern Australia) with data coverage above 50% (Sept. 2012 to Sept. 2013) and locations of the radar stations (RRK, NNB; black circles), ADCP moorings (CH70 and CH100; black squares) and Coffs Harbour weather station (black star). Color legend shows the standard deviation (STD, m s\(^{-1}\)) of the \(v\)-component of the HF radar currents (6 h smoothed) calculated from the mean variance for Sept. 2012 following methodology described in Wyatt et al. [2017]; (b) validation of surface HF radar currents against 10 m deep ADCP currents (6 h smoothed data; speed = \(\sqrt{u^2 + v^2}\)) at CH100 mooring shown by the linear regression (red line; equation in the figure; \(p < 0.01\)) and the joint probability density function normalized by its maximum value (PDF; color legend); (c) 6 h smoothed meridional (\(V\)-component, solid black line) and zonal (\(U\)-component, dotted black line) wind velocity components (m s\(^{-1}\)) measured at Coffs Harbour weather station (black star in Fig. 1a) every 30 minutes; durations of the frontal jet (blue line), cold-core eddy (CCE; orange line) and warm-core eddy (WCE; green line) events are highlighted (see legend on the top).
Figure 2: (a) Wind rotary spectra (Sept. 2012 – Sept. 2013) for U- and V-components (6 h smoothed data from Coffs Harbour weather station); red bars show the 95% confidence intervals using chi-square distribution with 62 degrees of freedom; the multitaper spectra [Lilly, 2016] were averaged in the frequency domain over 32 multiple Slepian tapers; the main peaks at 6 h, 12 h and 24 h are highlighted on the top; (b) maximum cross-correlation coefficient ($\rho_{xy}$) between surface meridional currents (6 h smoothed data) and north-south ACCESS model wind velocity every 12 km during Sept. 2012, showing the higher influence of the wind forcing on the surface circulation near the coast; (c) cross-correlation ($\rho_{xy}$, cross-correlation coefficient) between 6 h smoothed surface meridional currents at CH100 mooring for HF radar (top 0.9 m) and ADCP (10 m depth) and north-south wind velocity (measured at Coffs Harbour station), see legend in the Figure; the cross-correlations are shown for two periods: (i) 1-19 Sept. 2012 when the main EAC branch was offshore the radar domain (solid lines; EAC offshore) and the short-term surface circulation was mainly driven by winds and (ii) 20-30 Sept. 2012 when the main EAC branch encroached upon the shore (dotted lines; EAC nearshore); cross-correlation coefficient and 95% confidence intervals were calculated following Emery and Thomson [2004] and Bendat and Piersol [2010]; (d) same as (c) for correlations between zonal currents and east-west winds.
Figure 3: Warm frontal jet intruding Coffs Harbour region on 10 Sept. 2012 visible on (a) sea surface temperature (°C, color legend) and surface HF currents (black arrows, m s⁻¹); (b) Rossby numbers \( Ro = \frac{\zeta}{|f|} \) (color legend); darker blue and orange lines indicate cyclonic \( (Ro < 0) \) and anticyclonic filaments \( (Ro > 0) \), respectively (c) surface divergence \( \frac{\gamma}{|f|} \) (color legend); red \( (\gamma/|f| < 0) \) and green \( (\gamma/|f| > 0) \) tones indicate convergent and divergent regions, respectively; (d) HF radar
wind direction (3 km resolution) using constant wind speeds for representation (black arrows) and 12 km resolution ACCESS model wind direction and intensity (m s$^{-1}$, red arrows; see speed scale on the top left); times for the HF radar and ACCESS winds snapshots are indicated by HF and A initials, respectively (e) cross-self section on 30° 08'S showing the opposite cyclonic (anticyclonic) vorticities on the cold (warm) sides of the jet; cross-section position is indicated by orange lines in Fig. 3a,b,c. Blue lines show the 50, 100, 200 and 500 m isobaths from land to ocean.
Figure 4: The CCE formation and decay; top panels (a, b, c and g, h, i) show HF radar wind direction (3 km resolution) using constant wind speeds for representation (black arrows) and the 12 km resolution ACCESS model wind direction and intensity (m s⁻¹, red arrows; see speed scale on the top left) and bottom panels (d, e, f and j, k, l) show the HF radar surface currents (m s⁻¹, black arrows) and Rossby number ($Ro = \zeta / |f|$; color legend); times for the HF radar and ACCESS wind snapshots are indicated by HF and A initials, respectively. Panels show: (a) strengthened southward winds; (b, c, g) winds reversed to north; (h, i) northward winds weakened; (d) northern dark blue region of strong cyclonic vorticity; (e) CCE
started to form; (f, j) cyclonic ageostrophic circulation developed on the eastern flank of the CCE; (k) CCE started to decay and (l) dissipated. Blue lines show the 50, 100, 200 and 500 m isobaths from land to ocean.
Figure 5: The cold-core eddy on 14 Sept. 2012 visible on (a) sea surface temperature (°C; color legend) and surface HF currents (black arrows, m s⁻¹) and (b) surface divergence (γ/|f|, color legend); red (γ/|f| < 0) and green (γ/|f| > 0) tones indicate convergent and divergent regions, respectively. Histograms of the HF surface current direction normalized by the maximum count (values between 0 and 1) to facilitate comparison for days with (c) dominant northward winds opposing the frontal jet and (d) dominant southward winds favoring the frontal jet; black bars show the distribution during the frontal jet and CCE events and blue bars during the whole Sept. 2012. Mean, median, standard deviation, skewness and kurtosis values are shown in the figures for Sept. 2012 (left side in blue) and the event days (right side in black).
Figure 6: (a) The $v$-component of ADCP subsurface currents (below 10 m depth) and (b) temperature along the water column at CH100 mooring between 10 and 25 September 2012. Black rectangles indicate the life span of the frontal jet, CCE and WCE.
Figure 7: The warm-core eddy formation and decay; sea surface temperature (°C, color legend) and surface HF currents (m s⁻¹, black arrows) on (a) Sept. 20, (b) Sept. 21 and (c) Sept. 23; (c) surface divergence (∇/|f|, color legend) on the (d) Sept. 20, (e) Sept. 21 and (f) Sept. 23; (g) cross-section showing the strong cyclonic vorticity (ζ < 0) on the cold side of the northward frontal jet and anticyclonic vorticity (ζ > 0) on the warm side on Sept. 21 (see also Fig. 8j); cross-section position is indicated by orange lines in Fig. 7b,e and Fig. 8j.
Figure 8: The WCE formation and decay; legend caption as for Fig. 4. Panels show: (a) weak northward winds; (b) weak southward winds; (c, g) moderate southward winds; (h, i) moderate northward winds; (d, e, f) WCE developing; (j) WCE fully developed; (k) WCE stretched; (l) WCE dissipated.
Figure 9. Snapshots of particle distribution after 18 hours of simulation using the semi-empirical particle tracking model for (a) the frontal jet, (b) the cold-core eddy (CCE) and (c) the warm-core eddy (WCE). Mean relative diffusivities ($K_R$, m$^2$ s$^{-1}$; color legend) after 18 hours, plotted at the initial particle position for the (d) frontal jet, (e) CCE and (f) WCE. Histograms of the mean relative diffusivities ($K_R$, m$^2$ s$^{-1}$) normalized by the maximum count (values between 0 and 1) to facilitate comparison for the (g) frontal jet, (h) CCE and (i) WCE; y-axis was restricted to -5 m$^2$ s$^{-1}$ to +10 m$^2$ s$^{-1}$ to facilitate visualization; maximum positive $K_R$ values of 12, 53 and 137 m$^2$ s$^{-1}$ were observed during the frontal jet, CCE and WCE events, respectively.
Figure 10: (a) Chlorophyll-α from MODIS image on Sept. 09 (03:20 UTC) showing the southward advection of high chlorophyll-α waters within the frontal jet; (b) the offshore entrainment of productive shelf waters by the cyclonic eddy on Sept. 14 (03:45 UTC) with final particles position (yellow dots) after 18 h of tracking on top; (c) oligotrophic EAC waters on the shelf two days after the WCE (Sept. 23, 03:35 UTC) but enhanced chlorophyll-α concentrations along the submesoscale fronts. Black square shows the HF radar domain and black dots the HF radar stations.