Observations of Surface-Wave-Current Interaction

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Abstract

Wave-current interactions can result in significant inhomogeneities of the ocean surface wave field, including modulation of the spectrum, wave breaking rates, and wave statistics. We present novel airborne observations from two experiments: 1) the High-Resolution air-sea interaction (HiRes) experiment, with measurements across an upwelling jet off the coast of Northern California, and 2) an experiment in the Gulf of Mexico with measurements of waves interacting with the Loop Current and associated eddies. The significant wave height and slope varies by up to 25% due to these interactions at both sites, whereas whitecap coverage varies by more than an order of magnitude. Whitecap coverage is well correlated with spectral moments; negatively correlated with the directional spreading and positively correlated with the saturation, and saturation normalized by the directional spreading. Surface wave statistics measured in the Gulf of Mexico, including wave crest heights, and lengths of crests per unit surface area, show good agreement with second-order nonlinear approximations, except over a focal area. Similarly, distributions of wave heights are generally bounded by the generalized Boccotti distribution, except at focal regions where the wave height distribution reaches the Rayleigh distribution with a maximum wave height of 2.55 times the significant wave height; much larger than the standard classification for extreme waves. However, theoretical distributions of spatial statistics that account for second-order nonlinearities serve as an upper bound for the observed statistics of extreme wave elevations. The results are discussed in the context of improved models of breaking, and related air-sea fluxes.

1. Introduction

Surface wave processes have important applications in air-sea interaction, coastal circulation, ocean remote sensing, and offshore engineering. Surface waves are important for air-sea interaction, modulating the exchange of energy, momentum, heat, and mass between the ocean and the atmosphere. Wave breaking drives upper ocean currents and mixing (Phillips 1977), affects aerosol production (Lenain & Melville 2016), and enhances gas exchange across the air-sea interface (Thorpe 1982; Farmer et al. 1993), all of which have implications for climate change predictions (Loewen 2002).

Wave-current interactions, that is, the effects of currents on waves, include refraction due to wave propagation over spatially varying currents and wave action conservation which can result in wave steepening for waves encountering an opposing current and vice versa. Wave-current interactions modulate the ocean roughness (Phillips 1984; Munk et al. 2000) and can enhance nonlinear effects such as wave breaking, which affect satellite remote sensing products such as ocean color (Gordon 1997; Moore et al. 2000), and radar imaging (Phillips 1984; Kudryavtsev et al. 2005). Similarly, wave-current interactions can lead to the formation of extreme wave heights (White and Fornberg 1998, Onorato et al. 2011, Toffoli et al. 2015, Janssen and Herbers 2009). Wave breaking can be a good visual indicator of wave-current interactions. As shown by Melville et al. (2005), areas of enhanced wave breaking due to wave-current interaction are often associated with sea surface temperature (SST) fronts. An example of a feature studied in this paper is shown in

Figure 1 displaying a photograph of the sea surface with a "line" of enhanced breaking due to wave-current interaction. Surface waves can also be modulated due to changes in relative wind forcing and changes in stability of the atmospheric boundary layer across mesoscale oceanic fronts with warmer water leading to intensification of the surface winds (Friehe et al. 1991, Jury 1994), which in turn modulates the surface wave field (e.g. Hwang 2005), resulting in increased wave breaking (Chelton et al. 2006). Gallet and Young (2014) recently showed that wave refraction induced by the vorticity of the Antarctic Circumpolar and Equatorial surface currents can result in significant deviations of swell from great circle paths across the Pacific Ocean. This roughly explains the outliers from the analysis by Munk et al. (1963, 2013) who, by not accounting for refraction by currents, traced swell measured off Southern California along great circle paths back to Antarctica, rather than the South Pacific Ocean.

Several field studies have characterized the modulation of the wave field by tidal (Vincent 1979, Masson 1996, Pearman et al. 2014) and large-scale currents (Kudryavtsev et al. 1995, Wang 1994, Haus 2007). However, more measurements are needed to improve numerical wave models and wave-breaking parameterizations in conditions with strong wave-current interactions (e.g., Romero and Melville 2010a, Banner and Morison 2010, Ardhuin et al. 2010, 2012). This study presents novel wave observations from two experiments over areas with significant wave-current interactions, including characterization of the modulation of the directional spectrum, wave breaking and wave statistics. The paper is organized as follows: Section 2 introduces the environmental conditions, field experiments, and instrumentation; Section 3 describes the analysis and results, which are discussed and summarized in Sections 4 and 5, respectively.

2. Field Experiments

This study presents field observations from two experiments: 1) the Office of Naval Research (ONR) High-Resolution air-sea interaction DRI program (HiRes) and 2) an experiment in the northern Gulf of Mexico (GoM). Both field campaigns collected wave observations from aircraft allowing characterization of inhomogeneities of the wave field over areas with strong surface currents and current gradients. The HiRes data analyzed in this study were collected at the edge of an upwelling jet off the coast of Northern California where significant gradients of wave breaking were observed (Fig. 1). The experiment in the GoM gathered observations near the northernmost edge of the Loop Current, including an area of opposing waves and currents, and a focal region.

2.1. HiRes

The HiRes program was designed to study air-sea interaction processes with a focus on wave-phase-resolved physical processes, such as air-flow over waves including both realistic large-eddy simulations (Sullivan et al. 2014) and field observations (Grare et al. 2013). Other field measurements included broadband spectral distributions of breaking waves, from air-entraining to micro breakers (Sutherland and Melville 2013), surface turbulence measurements (Sutherland and Melville 2015), and aerosols (Lenain and Melville, 2016). This paper presents airborne wave observations over areas with significant wave-current interactions collected near

Bodega Bay on June 17, 2010, during upwelling conditions with 13 m/s winds from the northwest. A composite map of SST from MODIS data is shown in Figure 2a with the wind direction shown with a red arrow pointing towards the southeast. An upwelling front can be observed at about 50 km from the coast. Supporting surface current observations were available from an existing array consisting of 12 MHz High Frequency (HF) radars located near Bodega Bay and Point Reyes (Kaplan et al. 2005; Kaplan and Largier 2006). A snapshot of the surface currents measured by the coastal HF radars is shown in Figure 2b. The current map shows a strong upwelling jet, reaching maximum currents of 1 m/s, with strong horizontal shear on the eastern side where the area of enhanced breaking (Fig. 1) was observed, as shown by a solid white line. This study focuses on the modulation of the wave field across the edge of the coastal jet.

2.11 HiRes instrumentation

Several platforms operated during HiRes, including R/P Flip, the CIRPAS Twin Otter (TO) aircraft, R/V Sproul, and Aspen Helicopters' Partenavia P68-C aircraft. This study focuses on the measurements collected from the TO aircraft. The flight track is shown with gray lines in Figure 2. The TO was equipped with two downward looking lidars, the NASA/EGG scanning Airborne Topographic Mapper (ATM) and a fixed RIEGL (model LD90–3800EHS- FLP) nadir-looking lidar. The TO was also equipped with nadir visible and infrared (IR) imagers, a pressure sensor array in the nose cone to measure atmospheric turbulence and fluxes, a Heitronic KT19 radiometer for local SST measurement, and an aerosol measurement package. An inertial motion unit (IMU, Northrop Grumman LN100-G) with global positioning system (GPS, Applanix POS AV 510) was used for data georeferencing. A set of two nadir-looking IMPERX IPX-11M5-L (12bit, 4000x2672px) synchronized cameras, each sampled at 5Hz, were used to compute the breaking statistics described in section 3.12. Typical spatial resolution was approximately 10-15 cm depending on flight altitude, leading to an image width in the cross-track direction ranging from 250 to 400 m, and 400 to 600 m along-track. Sun glitter was minimized with electro-mechanically controlled linear polarizers.

The ATM is a conical scanning lidar used previously to measure directional wavenumber spectra of surface waves (Hwang et al. 2000a,b; Romero and Melville 2010b). During HiRes, the ATM had a pulse repetition rate of 5 kHz, a scanning rate of 20 Hz, and a conical scanning angle of 30^o. Thus for a nominal aircraft speed of 50 m/s at an altitude h= 300 m above mean sea level (AMSL), the theoretical horizontal resolution is δx=2.5 m along-track and δy=4.75 m cross-track, with a swath width SW of approximately 300 m (SW=2 h tan[30^o] ≈ h). The calibrated elevation error per pulse is approximately 8 cm (Krabill and Martin 1987). During post-processing the raw georeferenced ATM data were separated into forward and aft parts of the scan, then spatially binned on a regular grid with a resolution of 2.5 m, and empty cells were interpolated with MATLAB's TriScatteredInterp function. Thus, the highest wavenumber resolved k_h=1.25 rad/m. The lidar pulse return rate was 30±3% for the forward scan and 50±4%, for the aft scan due to the aircraft's angle of attack (3.6^o ± 0.5^o). The analysis of the data was done exclusively on the aft scan with the higher pulse return rate.

2.12 HiRes spectral analysis

Directional wavenumber spectra were calculated from the spatially interpolated ATM data using the Fast Fourier Transform and the squared amplitude of the Fourier coefficients. The directional wavenumber spectrum $F(k_x,k_y)$ is defined according to

$$\langle\eta^2\rangle = \int_{-k_h}^{k_h}\int_{-k_h}^{k_h} F(k_x,k_y)\, dk_x dk_y, \qquad (1)$$

such that its integral over all wavenumbers corresponds to the variance of the sea surface elevation.

Prior to calculating each spectrum, the edges of the data were tapered with a 10% Tukey window in two dimensions. Zero padding was applied in the cross-track direction, doubling the width to 600 m. Spectra were calculated from swaths 2.5 km long with 50% overlap, allowing the characterization of spatial inhomogeneities of the wave field. In addition to correcting the spectrum for the variance loss due to tapering, spectra were corrected for the Doppler shift induced by the relative motion between waves and the aircraft (Plant et al. 2005). Neighboring wavenumbers were averaged together in the along-track direction (x) yielding spectra with resolution Δk_x=Δk_y=2 π/600 = 0.01 rad/m. Resulting spectra were smoothed with a 3x3 top-hat filter, yielding 38 (2×300/600×2500/600×9) degrees of freedom (DOF) per spectrum. Finally, spectral energy densities at angles larger than ±90° from the wind direction were set to zero and the remaining spectral components were multiplied by two, preserving the variance with the oceanographic convention of energy propagating towards a given angle within the spectrum.

2.2 Experiment in the Gulf of Mexico (GoM)

The experiment in the GoM was designed to collect airborne observations of surface waves interacting with the Loop Current and related eddies in October 2011. It was conducted in the northern part of the Gulf when the Loop Current boundary was located very far north, overlapping in time when cold fronts are common in the GoM. These cold fronts generally propagate from Texas into the Northern Gulf during the fall and winter months (Henry 1979) giving rise to southward (northerly) winds followed by southwestward (northeasterly) winds as the fronts pass through. This allowed the possibility of investigating locally generated waves interacting with the Loop Current and eddies. This study focuses on the measurements collected on Oct. 30th, 2011, a day after the passage of a cold front, in winds of 8 m/s. Figure 3a shows SST analysis from the HYbrid Coordinate Ocean Model (HYCOM) at 1/25^{o} horizontal resolution in the northern Gulf of Mexico over the edge of the Loop Current. The corresponding surface currents are shown in Figure 3b. The red arrow indicates the mean wind direction towards the southwest, and the dashed box shows the study area.

2.21 GoM INSTRUMENTATION

The research platform was the Partenavia P68-C aircraft operated by Aspen Helicopters. Figure 3 shows the flight track with a thick black line going across the edge of the Loop Current. The aircraft was equipped with the Modular Aerial

Sensing System (MASS, Reineman et al. 2009, Melville et al. 2016, Clark et al. 2014) composed of a downward looking raster scanning lidar (Riegl LMS-Q680i), a long-wave infrared camera (QWIP FLIR SC6000), high-resolution video (JaiPulnix AB-800CL), and a hyperspectral imaging system (Specim EagleAISA). All instruments are time synchronized and georeferenced with a high-accuracy coupled GPS/IMU (Novatel SPAN-LN200). The raster lidar provides higher spatial resolution than the ATM and other airborne lidars used previously to measure ocean waves (c.f. Hwang et al. 2000a, Romero and Melville 2010b, Reineman et al. 2009; Huang et al. 2012).

The raster scanning lidar has a field of view (FOV) of 60^{o}, therefore the swath width on the ocean surface is approximately equal to the flight altitude h. The lidar operated mainly in two modes, each setting designed to capture different scales.

Mode 1: h= 200 m, sampling frequency f_s=266 kHz, scanning frequency f_{sc}= 200 Hz, and angular scan increment $\Delta\theta$=0.045^{o}, where $\Delta\theta$=FOV/M with M=f_s/f_c corresponding to the number of pulses per scan.

Mode 2: h= 550 m, f_s= 60 kHz, f_{sc}= 70 Hz, and $\Delta\theta$=0.07^{o}.

The nominal aircraft speed V_a was 50 m/s, thus the along-track resolution for Mode 1 is 25 cm (V_a/f_{sc}) and 71 cm for Mode 2. The cross-track resolution at nadir is 16 cm and 70 cm, which increases to 20 cm and 84 cm at the edge of the swath for Modes 1 and 2, respectively (Reineman et al. 2009). Following the method used to post-process the ATM data, lidar data were binned and interpolated on a regular grid with horizontal resolution of 0.5 (k_h= 6.28 rad/m) and 1.5 m (k_h=2.1 rad/m) for Modes 1 and 2, respectively, before computing smoothed directional wavenumber spectra from 5 km-long swaths with 50% overlap for 75 degrees of freedom, and a spectral resolution identical to that of the ATM ($dk_y = dk_x = 2\pi/600$ m = 0.01 rad/m).

2.22 GoM SPECTRAL ANALYSIS

A sample directional wavenumber spectrum from lidar measurements collected at 550 m AMSL in the GoM at the edge of the Loop Current is shown in Figure 4. The peak wavenumber k_p= 0.08 rad/m. The spectrum $F(k,\theta')$ is rotated such that $\theta'=\theta-\theta_w=0^{\circ}$ corresponds to the wind direction, with the wind blowing towards θ_w. The corresponding omnidirectional spectrum $\phi(k) = \int_{-\pi/2}^{\pi/2} F(k,\theta')k d\theta'$ is shown in Figure 5a and is compared with a spatially overlapped spectrum measured at a lower altitude (h= 200 m). Both spectra agree for $k < 1$ rad/m, approximately following a $k^{-2.5}$ power law. At larger wavenumbers the spectral tail can be better approximated by a k^{-3} power-law. The corresponding saturation spectra $B(k) = \phi k^3$ are shown in Figure 5b. The directional spreading $\sigma(k)$ is defined as the root-mean-square directional width from the mean spectral direction $\bar{\theta}$ according to

$$\bar{\theta}(k) = \frac{\int_{-\pi/2}^{\pi/2} F(k,\theta')\theta' d\theta'}{\int_{-\pi/2}^{\pi/2} F(k,\theta') d\theta'}$$

and

$$\sigma(k) = \left(\frac{\int_{-\pi/2}^{\pi/2} F(k,\theta')[\theta' - \bar{\theta}(k)]^2 d\theta'}{\int_{-\pi/2}^{\pi/2} F(k,\theta') d\theta'} \right)^{1/2}. \qquad (2)$$

The spreading is narrowest near the spectral peak ($\sim 15^{\circ}$) increasing towards both low and high wavenumbers, approaching 55° at wavenumbers much larger than the peak (Fig. 5c). Both directional spreading curves are in good agreement except near and below the spectral peak, where the spectrum calculated from a wider swath is narrower (blue curve), and more accurate as it is better able to resolve the directionality of the lower wavenumbers. The normalized saturation defined by $\tilde{B}(k) = B(k)/\sigma(k)$ is an important parameter for the characterization of wave breaking (Banner et al. 2002; Romero et al. 2012) and is shown in Figure 5d.[2] When compared to the saturation spectrum, the normalized saturation exhibits a weaker increasing trend with increasing wavenumber.

3. Results

3.1. HiRes

During the HiRes program, on June 17th 2010, field observations were collected off Bodega Bay during upwelling conditions with steady 13 m/s winds to the southeast and waves near full development with significant wave height of 3 m. Measurements were collected over areas with strong wave-current interaction detected visually from aircraft by enhanced breaking and remote sensing of SST gradients near the edge of an upwelling jet. Figure 1 is a handheld photograph looking north, showing an area of enhanced breaking located at the edge of the jet (see also Fig. 2b). Also apparent are significant differences in the whitecap coverage on the left (west) side of the photo when compared to the right (east). A snapshot of SST measured by the IR imager on the Twin Otter is shown in Figure 6a. The data were collected over the area of enhanced wave breaking and it shows sharp SST gradients of about 0.4 degrees over O(10) m, corresponding to a submesoscale front partially aligned with

[2] Note that in Figure 5c, the directional spreading, σ, is given in degrees for convenience, whereas in the normalized saturation, B/σ it is given in radians.

the area of enhanced breaking. The dashed box on the bottom corresponds to the zoomed-in area of Figure 6b, which shows frontal instabilities with horizontal scales of the order of 100 m, comparable to the wavelength of the dominant waves. The corresponding georeferenced visible imagery from the downward looking camera is shown in Figure 6c, with the yellow dashed line showing the approximate location of the submesoscale front. The whitecap coverage is substantially different on each side of the front, with little or no breaking to the left (west). The photos in Figures 1 and Figure 6c were taken from different aircraft. The area of enhanced breaking from Figure 1 is not evident in Figure 6c, which only shows increased wave breaking on the lower right corner. This is likely due to differences in dynamic range, lighting and field of view between the downward looking and handheld cameras. The handheld photo taken at a grazing angle covers a much larger surface area and therefore it is easier to identify the coherent line of breaking. The position of the "line" of enhanced breaking was determined visually from the aircraft (white line in Figures 2a,b) within $\pm$ 1 km as it moved steadily towards the west. The line of breaking was located over an area with strong current gradients, with current speeds of 0.8 m/s to the west and decreasing down to about 0.4 m/s to the east (Fig. 2b).

A spatial cross section of the submesoscale front as measured by the Twin Otter aircraft is shown in Figure 7a, where x=0 is located within the core of the upwelling jet. The temperature drops by about 0.4 degrees between x=7 and x= 8 km. The vertical vorticity calculated from HF current data in Figure 7b, shows a maximum value at the edge of the temperature front. The whitecap coverage as measured from the airborne visible imagery (Fig. 7c) shows little breaking for 1 km < x < 7 km followed by increased breaking for x >7 km, especially at the edge of the front. A corresponding atmospheric response can be observed in the mean winds as measured from the aircraft at 30 m AMSL (Fig. 7d). The mean wind speed decreases over the segment with increased whitecap coverage to the right of the front (7 km < x < 10 km), illustrating the coupling between the atmosphere, ocean waves and upper ocean currents at horizontal scales of the order of 1 km.

*3.*11 WAVE FIELD MODULATION

In contrast to single point measurements, airborne lidar measurements allow analysis of the spatial inhomogeneities of the wave field due to wave-current interactions. Here, data collected along the flight track is analyzed using an objective analysis which maps randomly spaced data on a specified set of locations using weighted averages that depend on the spatial covariance of the data (Bretherton et al, 1976; Davis 1985). Following Denman and Freeland (1985), the structure function $G(r)$ of the significant wave height $H_s=4\,\langle\eta^2\rangle^{1/2}$ was calculated from data pairs and fitted to the model $G(r) = 2V[\varepsilon + 1 - H(r)]$, where r is the distance between pairs of measurements, V is the variance, ε is the fraction of noise variance, and $H(r) = \exp\left(\frac{-r^2}{2L^2}\right)$ is the assumed Gaussian autocorrelation function. The fitted noise variance ε = 0.1 and decorrelation length L= 2.8 km, which is comparable to the spatial resolution of the data (2.5 km) or two adjacent data points with 50% overlap (2 x1.25 km). Both parameters were used to calculate the objective map of H_s shown in Figure 7e. The significant wave height, H_s, is on average 2.8 m, varying within

25% over the area covered by the aircraft. It is generally lower over the southeast corner of the figure, particularly to the right (east) of the jet and increases by about 8% towards the jet core, from right to left (east to west). The peak wavelength is less affected with a mean value of 112 $\pm$ 14 m, corresponding to 0.056 $\pm$ 0.007 rad/m. This suggests that the modulation of the wave field across the coastal jet by the currents is mostly confined to wavenumbers larger than the spectral peak.

The airborne visible imagery collected from the downward looking camera allowed the quantification of the spectral statistics of breaking fronts, specifically $\Lambda(c_b)dc_b$, defined as the average length of breaking crests with speed in the range c_b to $c_b+ dc_b$ per unit surface area (Phillips 1985). Following Kleiss and Melville (2010, 2011), $\Lambda(c_b)$ was calculated from visible imagery (ImperX IPX-11M5-L dual camera system) collected on both sides of the submesoscale front (or "line" of breaking), which is shown in Figure 8 with the red and blue lines corresponding to the warm (west) and cold (east) side of the front, with sampling locations shown with red and blue lines in Figure 7e. Each collected digital image was first georeferenced using the information from the onboard GPS/IMU, then interpolated to a regular grid with a 10 cm spatial resolution. A detailed description of the breaking statistics processing steps used in the present study is provided in Kleiss and Melville 2011. The black dashed line is a reference power-law of c_b^{-6} according to Phillips' (1985) equilibrium model. The $\Lambda(c_b)$ distributions exhibit substantial variability, with larger values on the cold side of the front for 0.4< c_b < 2 m/s and 4 m/s < c_b < 10 m/s. Note that for c_b <2-3 m/s, the lack of air entrainment in the sampled breakers leads to a roll off of the distribution from the c_b^{-6}, as the visible imagery cannot accurately capture the breaking fronts (Romero et al. 2012). This is consistent with Sutherland and Melville (2013) who, by using a combination of infrared and visible cameras, showed that the c_b^{-6} behavior extended to much lower values of c_b (0.1-0.8 m/s). The inset shows $\Lambda(c_b)$ compensated by c_b^6, varying by up to a factor of 5 between the warm and cold areas. The speed of the breaking front, c_b, is linearly related to the wave phase speed c through a proportionality factor α such that $c_b=\alpha c$, with α varying within 0.7 and 0.9 (Rapp and Melville 1990; Stansell and MacFarlane 2002; Banner and Peirson 2007; see also Banner et al. 2014 and Pizzo & Melville 2015). From the linear dispersion relationship assuming $\alpha=0.8$ (Rapp and Melville 1990), the range of breaking speeds $4 < c_b < 10$ m/s, where the $\Lambda(c_b)$ distributions vary the most, corresponds to wavenumbers in the range of 0.06 to 0.4 rad/m (±25%).

We further examined the spatial modulation of the directional wavenumber spectrum focusing on the measurements around the area of enhanced wave breaking. Objective maps of mean saturation, directional spreading and normalized saturation were calculated using the noise variance and decorrelation length obtained from the structure function of the significant wave height, as shown in Figures 9a,b, and c, respectively. The spectral moments were averaged for $k_p \leq k <$ 0.4 rad/m, which is consistent with the range of wavenumbers where $\Lambda(c_b)$ varied the most across the front. All three parameters show substantial variability across the “line” of breaking when compared to H_s (Fig. 7e). The mean saturation increases, the spreading decreases, and the normalized saturation increases from left to right (west to east) across the “line” of breaking.

The spatial inhomogeneities of the wave field across the front are further analyzed along the sampling tracks with the available overlapping whitecap coverage, W, and directional wavenumber spectra. Figure 10a shows a spatial scatter plot color coded by W. There is substantial variability in W across the front, with very low values just to the left (west) of the front, and relatively larger values to the right (east). The area identified as the "line" of enhanced breaking, partially overlapping with the submesoscale front, is apparent with large values of whitecap coverage. The corresponding mean saturation and normalized saturation averaged in the range $k_p < k < 0.4$ rad/m are shown in Figures 10b,c, respectively. There is good spatial correspondence between W and $\langle B \rangle$, and the mean normalized saturation $\langle \tilde{B} \rangle$, with low values to the left (west) of the front, and relatively larger values to the right (east). Correlation coefficients R between mean spectral moments and W are significant with R= 0.64, -0.59, and 0.80 for the saturation, directional spreading, and normalized saturation, respectively, with the normalized saturation giving the best correlation. However, the correlation difference between W and saturation $\langle B \rangle$, and the normalized saturation $\langle \tilde{B} \rangle$ is not statistically significant with a p-value of 0.1 for a two-sided test. Extending the wave number range to $k_p < k < 1.0$ rad/m to compute the average moments, the resulting correlation coefficients are slightly reduced to R= 0.46, -0.43, and 0.67 for the saturation, directional spreading, and normalized saturation, respectively.

3.12 RAY TRACING

A ray tracing analysis was carried out following Mathiesen (1987), using the approximation that the local curvature of a ray is given by the vorticity field $\zeta = v_x - u_y$ divided by the group velocity c_g (Kenyon 1971; Dysthe 2001). The ray equations were integrated using the HF radar currents, and the measured mean peak wavenumber. The resulting rays are shown in black in Figure 11, plotted over the vorticity field normalized by the Coriolis parameter, f. The rays are parallel or divergent immediately to the west of the front over the sampling area. On the right side of the front (east) the rays are generally convergent over the sampling area, except over the southernmost part. This is consistent with the right-left (east-west) asymmetry of the whitecap coverage (Fig. 1, Fig 7c). It is also consistent with the wider spectrum to the left (west) and relatively narrower spectrum to the right (east) of the front (Fig. 9b). There is also ray convergence farther to the left of the "line" of breaking (west), where H_s and the saturation are also large: Fig. 7e and 9a, respectively. The spatial overlap between the line of enhanced breaking and the area of maximum current gradient (vorticity) suggests that enhanced breaking was likely a result of opposing waves and currents due to waves leaving the jet encountering an "opposing" current in a frame of reference relative to the jet. Repeating the ray tracing computations for wavenumbers within the range of increased values of $\Lambda(c_b)$, for example 4 k_p, results in qualitatively similar ray patterns but with enhanced ray curvature (not shown).

3.2. Experiment in the Gulf of Mexico

In the fall of 2011 the Loop Current extended very far north in the GoM, within range of the Partenavia aircraft based at Gulf Shores, Alabama. This, combined with the high probability of offshore winds due to frequent atmospheric cold fronts during that time of the year, provided an opportunity to collect airborne measurements of waves interacting with the Loop Current (LC) and related eddies. On October 30th, the Partenavia aircraft collected measurements near and across the edge of the Loop Current after the passage of an atmospheric cold front.

As described above, the structure function was calculated using data pairs of significant wave height and used to fit a decorrelation length and fractional noise variance assuming a Gaussian decorrelation function, yielding L= 9 km and ε = 0.07. These parameters were used to generate objective maps of H_s and other variables.

Figure 12a shows an objective map of H_s and dominant wave direction θ_p (black arrows). The gray arrows show the surface currents from HYCOM analysis. The edge of the LC is on the lower right corner of the figure. There is substantial variability of the dominant waves with H_s varying by as much as 25% (between 1.35 and 1.75 m) and θ_p varying by about 45^o. The dominant waves propagate to the west at the top of the figure and towards the southwest near the LC edge. There are three maxima in H_s, one over the area of opposing waves and currents, and the second to the north over a region of current convergence with collinear waves and currents, and the third in the NW corner. The objective map of significant wave slope $\eta_{rms}\, k_p$, defined as the product of peak wavenumber times the root-mean-square surface elevation $\eta_{rms}= \langle \eta^2 \rangle^{1/2}$ shown in Figure 12b, has a distribution similar to H_s but with values consistently larger over the area of opposing waves and currents in the LC. The whitecap coverage, W, in Figure 12c only shows two maxima, one maximum in the LC, and the second on the NE corner adjacent to an area of large H_s.

Objective maps of mean saturation, spreading, and normalized saturation are shown in Figures 12d,e, and f, respectively. All three parameters were averaged for $k_p< k <$ 1 rad/m, with the upper limit just before the noise floor of the spectra measured at the higher altitude (see Figure 5). The saturation is large over the area of opposing waves and currents, consistent with the significant slope, but exhibits another maximum on the northern-most part of the mapping area. The directional spreading varies by up to 8 degrees, showing an east–west asymmetry. The variability of the mean normalized saturation is qualitatively similar to the mean saturation but shows better correspondence with W. The correlations between W and $\langle B\rangle$, $\langle\sigma\rangle$, and $\langle\tilde{B}\rangle$ along at the sampling locations along the flight track give R = 0.45, -0.34 and, 0.54, respectively, with $\langle\tilde{B}\rangle$ giving the best correlation, consistent with the HiRes observations, but again the correlation difference between $\langle\tilde{B}\rangle$ and $\langle B\rangle$ against W is not statistically significant.

3.21 Wave statistics

Here we first investigate the modulation of statistical distributions of wave heights, crests, troughs, and crest-length due to wave-current interactions. The results are compared with analytical models including linear and nonlinear approximations. For this analysis the wave data were divided into three subsets: S1, S2, and S3, which are shown in Figure 12b delineated with dashed black lines. The groups where chosen based on similarities of significant slope with nearby wave observations. S1, S2, and S3 contain 25, 55, and 26 data swaths, with mean significant slope $\eta_{rms}\, k_p$ = 0.032±0.003, 0.028±0.003, and 0.036±0.005, respectively. S3 has the largest average wave slope and S2 the lowest, consistent with Figure 12b.

Individual crest and trough heights were calculated from each data swath along parallel lines in the direction of the dominant waves. Crests and troughs heights were calculated from the maximum and minimum elevation, respectively, between successive upward zero-crossings. Individual wave heights were determined from the difference between crest and trough heights. Probability density functions (pdf's) of crests η_c heights, magnitude of trough heights $|\eta_t|$, and wave heights H normalized by the root-mean-square of surface elevation η_{rms} where calculated for each data swath. Then, pdf's from multiple swaths were ensemble averaged within each data group (S1-S3). The exceedance probability of η_c/η_{rms} and $|\eta_t|/\eta_{rms}$ are shown in Figures 13a,b, respectively. The subsets S1, S2, and S3 are shown with red, green, and blue lines, respectively, with horizontal bars corresponding to the uncertainty due to standard error of η_{rms}, defined as 2 std/ $\sqrt{N}$, where std is the standard deviation and N is the number of samples within each data subset.

The wave crest distributions are bounded by the nonlinear Tayfun distribution with parametric dependence on the significant slope (Tayfun 1980; Toffoli et al. 2008), except for S1, where significant deviations from second-order theory can be observed. The wave trough distributions are generally lower than the Rayleigh distribution, in good agreement with Tayfun's distribution. The exceedance probability of wave heights, H, normalized by the significant wave height, H_s= 4 η_{rms} , is shown in Figure 13c and compared to the generalized Boccotti (GB) distribution, which includes effects due to finite spectral bandwidth and third-order nonlinear corrections (Alkhalidi and Tayfun, 2013). The GB distribution was calculated using the spatial two-dimensional autocorrelation function (Romero and Melville 2011) and the fourth-order cumulants of the sea surface elevation. It generally bounds the lidar measurements, except for S1 approaching the Rayleigh distribution with the maximum wave height approaching 2.55 H_s. This is much larger than the typical threshold for an extreme wave of 2 H_s.

Following the work by Romero and Melville (2011) the statistics of crest lengths were analyzed by defining $L\eta_o$ as the length of crests per unit surface area exceeding elevation threshold η_o. For this analysis the three-dimensional lidar data were thresholded at several values of positive surface displacement and then binarized. The binary images were then used to determine the length and orientation of each thresholded crest by fitting an ellipse. This allows us to calculate $L\eta_o$ and compare it against analytical distributions, including both linear and second-order nonlinear approximations, derived by Romero and Melville (2011) based on the statistical analysis of a random moving surface by Longuet-Higgins (1957). Figure 13d shows ensemble averages of $L\eta_o$ plotted against $(\eta_o/\eta_{rms})^2$. The measured distributions are well approximated by the nonlinear distribution by Romero and Melville (2011) shown with triangles, except for S1 which shows significant deviations from the second-order nonlinear distribution for large wave elevations.

The analytical distributions shown in Figures 13a-c are based on wave models at a single point. However extreme wave statistics at single point differ substantially from spatial and spatio-temporal statistics, with the latter giving the largest expected waves as the total number of waves increases (Dysthe et al. 2008; Fedele 2012). Recent studies have shown that the theoretical models of space-time statistics of extreme waves that account for second-order nonlinearities are consistent with spatio-temporal measurements collected in the Mediterranean Sea (Fedele et al. 2013; Benetazzo et al. 2015). The modeling study by Barbariol et al. (2015) suggests that space-time distributions of extreme wave heights normalized by η_{rms} increase only slightly by a few percent over areas of opposing currents due to modulation of the spectrum by currents. Here we compare our measurement of the extreme wave elevations against theoretical distributions of spatial extremes. Following Fedele et al. 2013 and Benetazzo et al. 2015, the exceedance probability of wave extremes for directional spectrum of random linear waves over a given area can be approximated by

$$P(\eta/\eta_{rms} > \tilde{\eta}) \approx 1 - (1 - \tilde{\eta}\ \exp[-x\ \tilde{\eta}/2])^{N_s}, \qquad (3)$$

where η_{max} is defined as the maximum surface elevation max[$\eta(\boldsymbol{x})$] within a spatial ensemble with N_s number of waves[3], which is proportional to the sampling area divided by the product of the mean wavelength times the mean crest-length (see appendix). Accounting for second-order bound harmonics the nonlinear surface elevation

$$\hat{\eta} = \tilde{\eta} + \frac{\mu\,\tilde{\eta}^2}{2} \rightarrow \tilde{\eta} = \frac{-1+\sqrt{1+2\mu\hat{\eta}}}{\mu}, \qquad (4)$$

where μ is a measure of the wave steepness accounting for a correction due to spectral bandwidth (also defined in the appendix). The probability of extreme surface elevations for nonlinear waves can be directly obtained from equations (3) and (4). The total number of waves N_s estimated for each data subset S1-S3 are:

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[3] Equation (3) does not include an additional term due to the number of the waves along the perimeter of the sampling area because it is negligible for relatively large areas.

2.9×10^4, 5.9×10^4, and 3.0×10^4, with mean steepness μ=0.045±0.002, 0.044±0.007, and 0.047±0.006, respectively.

The exceedance probabilities of extreme surface elevations were calculated combining all data within each subset (S1-S3) defining η_{max} as max[$\eta(\boldsymbol{x})$] over each 5 km long record. The measured distributions are shown in Figure 14 with red, green, and blue symbols corresponding to S1-S3, respectively. The corresponding linear and nonlinear theoretical distributions calculated from the average moments of the directional spectrum are shown with dashed and solid lines, respectively. The data generally exceed the linear model but are bounded by the nonlinear distributions, even in S1 (within error bars), where the largest waves are found. This suggests that theoretical distributions of 2nd order nonlinear space-time statistics of extreme waves are suitable for engineering applications even in conditions with strong wave-current interactions. However, the analytical model cannot explain the relative differences in observed extreme wave heights between the different data subsets (S1-S3).

3.22 Ray tracing

A ray tracing analysis was carried out using the observed mean peak wavenumber and direction computed over the sampled area and HYCOM surface current data. Figure 15 shows the resulting rays plotted over the vorticity field normalized by the Coriolis parameter. The white dots show the location of the wave measurements from the aircraft and the gray lines show the delineation between the different subsets (S1-S3). The rays show significant divergence over S2 where observed wave height was low (Fig 12a). In contrast the rays converge over S1 where H_s is largest.

Also, the focal area corresponds to the data subset where the normalized maximum wave heights (H/H_s), and extreme elevations (η_{max}/η_{rms}) are largest.

4. Discussion

The data from both the HiRes experiment off the coast of Northern California and the experiment in the Gulf of Mexico showed substantial inhomogeneities of the wave field due to wave-current interactions. In the context of wave breaking, wave-current interactions can have important implications for mixing and gas exchange between the ocean and the atmosphere. For example, the HiRes measurements showed enhanced wave breaking on the colder side of the submesoscale front. In the context of frontal dynamics, secondary circulation results in surface convergence at fronts (McWilliams 2016). This suggests that gas exchange may be enhanced not just due to enhanced wave breaking alone but also due to secondary ageostrophic circulation efficiently entraining bubbles down into the water column. Moreover, secondary circulation at fronts depends on vertical mixing (McWilliams et al. 2015), which can in turn be modulated by wave-current interactions asymmetrically across fronts.

Other possible important feedbacks include spatial gradients of the surface momentum flux due to modulation of wave breaking by wave-current interactions, and vortex forces due to shear-induced refraction (McWilliams et al. 2004; Sheres and Kenyon 2006) and related Langmuir circulation. The frontal instabilities shown in Figure 6a,b have scales comparable to the dominant wavelength of the surface waves, further suggesting the possibility that the separation of frontal and surface-wave scales may not generally apply.

As the various remote-sensing applications continue to evolve towards finer spatial resolutions, for example ocean color and altimeters, detailed knowledge of the surface wave field and its inhomogeneities due to wave-current interactions will become increasingly important. For both active and passive remote sensing, the fine-scale structure of the ocean surface is of fundamental importance, and the modulation of this structure will be affected by the wave-current interaction processes described here.

On the incidence of extreme waves, we found that the measurements were generally bounded by theoretical nonlinear distributions of spatial wave extremes. However, the theoretical model fails to describe the trend between focal and non-focal areas based on modulation of the measured spectrum. Our measurements without temporal information may under-sample extreme wave heights. However, higher order nonlinearities cannot be ruled out for the observed extreme waves within the focal area (e.g Janssen and Herbers 2009; Onorato et al. 2011; Toffoli et al. 2011, 2015).

Regarding the characterization of wave breaking with respect to the modulation of the spectrum, the data consistently gave larger correlation coefficients between the whitecapping coverage against the normalized saturation. Following a suggestion from an anonymous reviewer, we also tested an anisotropic spectral saturation metric introduced by Ardhuin et al. 2010[4], which is given by

$$\hat{B}(k,\theta) = \int_{\theta-\Delta_\theta}^{\theta+\Delta_\theta} F(k,\varphi)k^3\cos^2(\theta-\varphi)k\,d\varphi, \qquad (5)$$

where Δ_θ=80°. It was found that mean anisotropic saturation $\langle\hat{B}(\bar{\theta}_B)\rangle$ along the mean saturation direction $\bar{\theta}_B$ with

$$\bar{\theta}_B = \frac{\int\int F(\boldsymbol{k})\theta k^3 d\boldsymbol{k}}{\int\int F(\boldsymbol{k})k^3 d\boldsymbol{k}},$$

correlated the best with the whitecapping coverage. The correlation coefficients obtained are 0.71 and 0.56 for the HiRes and GoM datasets, which are similar to those obtained with the normalized saturation (i.e., 0.80 and 0.54, respectively). But again, the correlation differences are not statistically significant compared to those obtained using the mean saturation $\langle B\rangle$.

5. Conclusions

We have presented a characterization of inhomogeneities of the ocean surface wave field over areas with strong wave-current interactions. This was accomplished with novel airborne observations collected during HiRes near Bodega Bay and an experiment in the Gulf of Mexico. Both data sets showed modulation of the wave height due to wave-current interactions by 25%. The analysis from HiRes observations focused on measurements collected on the edge of an upwelling jet, where strong gradients of wave breaking were found. An area of enhanced breaking was identified at the edge of the jet, overlapping with a submesoscale front. The area

[4] Corrected without the factor of c_g $(2\pi)^{-1}$.

of enhanced wave breaking separated two breaking regimes, with little breaking to the west and relatively more breaking to the east over the colder SST. Measurements across the submesoscale front showed maximum vertical vorticity at the edge of the front, and a reduction of the mean winds at 30 m AMSL over the areas with larger whitecapping coverage, which is consistent with an increase of the drag coefficient due to increased wave breaking. Analysis of the wavenumber spectra across the jet showed that the mean saturation B, directional spreading σ, and normalized saturation $\tilde{B}$ varied substantially across the jet, correlating well with the whitecap coverage.

The measurements in the Gulf of Mexico were collected over the edge of the Loop Current and associated eddies after the passage of a cold front. The wave field showed substantial modulation due to currents, including conditions of opposing waves and currents and a focal area. The measured whitecap coverage correlated well with the spectral moments for wavenumbers larger than the spectral peak.

Statistical analysis of wave crests and length of crests per unit area showed agreement with analytical distributions from second-order nonlinear approximations, except over the focal area where significant deviations from second-order nonlinear theory were found. Similarly, measured wave height distributions were generally bounded by the generalized Boccotti distribution except over the focal area where the wave height distribution reached the Rayleigh distribution, with H_{max} = 2.55 H_s, which is much larger than 2H_s, the typical threshold criterion used to define extreme waves. However, the measured statistics of extreme wave elevations were bounded by analytical second-order nonlinear distributions of spatial extremes.

Finally, it is important to appreciate that surface wave measurements having the accuracy and spatio-temporal coverage displayed here would not have been possible without the advantages of airborne measurements; firstly, to find regions of strong wave-current interaction, and secondly, to be able to measure the wave fields over large areas with the accuracy described here.

APPENDIX A

Wave Parameters

Following Fedele et al. (2012, 2013), the number of waves over an area $L_x L_y$ is given by $N_s = \sqrt{2\pi}\frac{L_x L_y}{\bar{\lambda}_x \bar{\lambda}_y}\sqrt{1-\alpha^2}$, where L_x,L_y are the length and width of the wave record, the corresponding mean wavelengths $\bar{\lambda}_x = 2\pi\sqrt{\frac{m_{00}}{m_{20}}}$, $\bar{\lambda}_y = 2\pi\sqrt{\frac{m_{00}}{m_{02}}}$, and $\alpha = m_{11}/\sqrt{m_{02}m_{20}}$. The moments of the directional spectrum are given by $m_{ij} = \iint k_x^i\, k_y^j F(\boldsymbol{k}) d\boldsymbol{k}$. Although the steepness parameter μ is often defined as the product of $\eta_{rms}\, k_p$ (e.g., Mori and Janssen 2006; Romero and Melville 2011), for consistency with Fedele et al. (2013), here μ is defined from moments of the frequency spectrum according to $\mu = \frac{\eta_{rms}\,\bar{\omega}^2}{g}(1-\nu+\nu^2)$, where $\bar{\omega} = m_1/m_0$ is the spectrally weighted mean frequency, and $\nu = \sqrt{m_0 m_2/m_1^2 - 1}$ is a measure of the spectral bandwidth. The frequency spectrum $\Psi(\omega)=\phi(k)\partial k/\partial\omega$, with $\omega=(g\,k)^{1/2}$ according to the linear dispersion relationship, and the moments $m_i = \int \omega^i \psi(\omega) d\omega$.

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Fig. 9. Objectively mapped spectral moments across the area of enhanced wave breaking (thick black line) from lidar data. Degree of saturation $\langle B\rangle$ (a), directional spreading $\langle\sigma\rangle$ (b), and normalized saturation $\langle\tilde{B}\rangle$ (c). The brackets represent averages in the range $k_p \le k <$ 0.4 rad/m, where k_p=0.056 rad/m. The gray dots indicate the mean sampling locations by the lidar.

Fig. 10. (a) Fractional whitecap coverage W across the sea surface temperature front. (b,c) Mean saturation and normalized saturation, respectively, for k_p rad/m < k < 0.4 rad/m. The thick black line shows the location of the area of enhanced wave breaking.

Fig. 11. Ray tracing over the surface vorticity field of the HF radar current data. The vorticity ζ is normalized by the Coriolis parameter f. Ray trajectories were integrated from the NW with mean peak wavenumber k_p and direction computed over the sampling area. The thick gray line shows the location of the area with enhanced wave breaking. The white dots show the mean sampling locations by the lidar.

Fig. 12. Objective maps of (a) Significant wave height H_s, (b) significant slope $\eta_{rms} k_p$, (c) fractional whitecap coverage W, degree of saturation $\langle B\rangle$ (d), directional spreading $\langle\sigma\rangle$ (e), and normalized saturation $\langle\tilde{B}\rangle$(f), with the brackets corresponding to a spectral average for $k_p < k <$ 1.0 rad/m. The black arrows (a) show the dominant wave direction and the white dots (a,c) show the mean sampling locations. The gray vectors show surface currents from HYCOM analysis. The dashed black lines delineate the three data groups S1, S2, and S3 used to compute the wave statistics (Fig. 13).

Fig. 13. Measured statistical distributions from the data subsets S1, S2, and S3, corresponding to the red, green, and blue lines, respectively. Exceedance probability distributions of wave crests η_c and troughs $|\eta_t|$ normalized by the root-mean-square surface elevation η_{rms} are shown in panels (a) and (b), respectively. The triangles in (a,b) correspond to second order Tayfun distributions with parametric dependence on the significant slope. Panel (c) shows the exceedance probability of wave height H normalized by the significant wave height H_s. The colored dashed lines

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Fig. 14. Probability of exceedance of dimensionless extreme wave elevations η_{max}/η_{rms}. The circles, triangles, and squares correspond to data subsets S1, S2, and S3, respectively. The corresponding linear and nonlinear distributions of wave extremes over an area (Fedele et al. 2013) are shown with dashed and solid lines, respectively. The error bars are calculated from the error of the mean of η_{rms} due to spatial inhomogeneities.

Fig. 15. Ray tracing over the surface vorticity field of HYCOM 4 km surface current data in the northern Gulf of Mexico. The vertical vorticity $\zeta = v_x - u_y$ is normalized by the Coriolis parameter f. Ray trajectories (black lines) were integrated with measured mean peak wavenumber k_p and direction and assumed constant from the NE. The white dots show the location of the airborne wave measurements. The solid gray lines delineate the three data groups (S1-S3) used to compute the wave statistics (Figs. 13 and 14).

Fig. 1. Area of enhanced breaking due to wave-current interactions. Photo taken looking North off the coast of Bodega Bay, California, June 17, 2010. The black region in the lower right of the image is due to the frame of an open window of the aircraft.

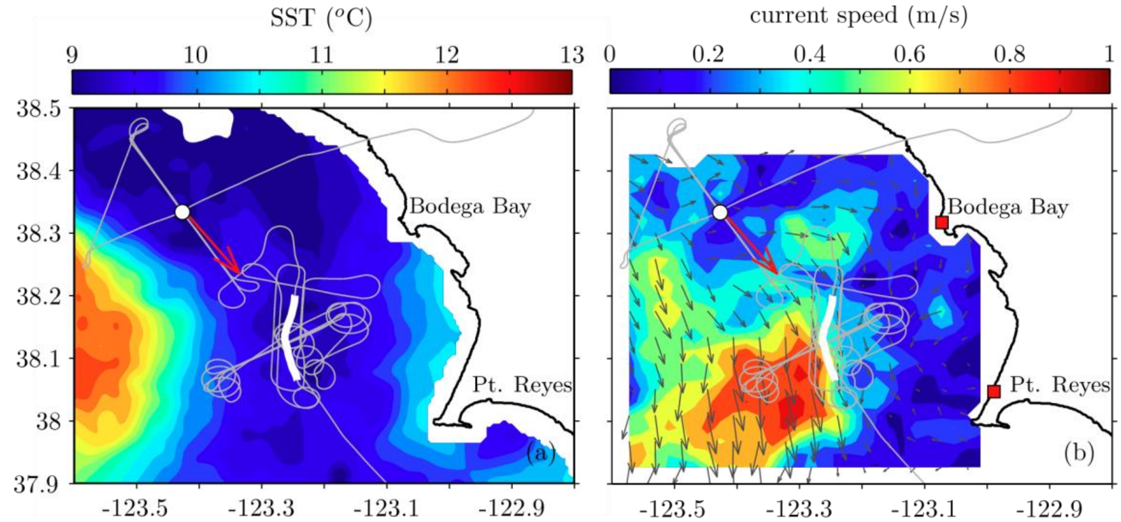


Fig. 2. Panel (a) shows a composite map of sea surface temperature off northern California from MODIS sensors onboard Aqua and Terra satellites on June 17, 2010. Panel (b) shows HF radar surface currents with vectors decimated by a factor of two. The gray line shows the fight track of the Twin Otter aircraft and the thick white line indicates the area of enhanced breaking (see Figure 1). The red arrow indicates the mean wind direction measured at FLIP (white circle). The red boxes in (b) show the location of the HF radars.

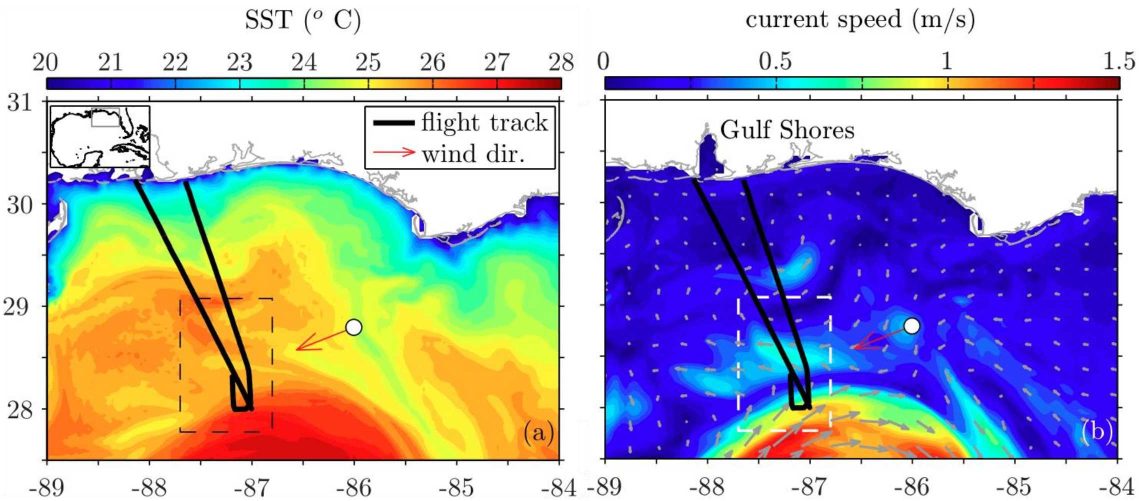


Fig. 3. Sea surface temperature (a) and surface current (b) from HYCOM reanalysis in the northern Gulf of Mexico on October 30, 2011. The gray line shows the coastline, the thick black line shows the flight track, and the red arrow indicates the mean wind direction from NDBC buoy 42039 (white circle). The dashed box indicates the study area. The gray arrows in (b) are the current vectors decimated by a factor of 7. The inset in panel (a) shows the coastline around the Gulf of Mexico and parts of the Caribbean Sea with the gray box indicating the corresponding zoomed-in area of the figures.

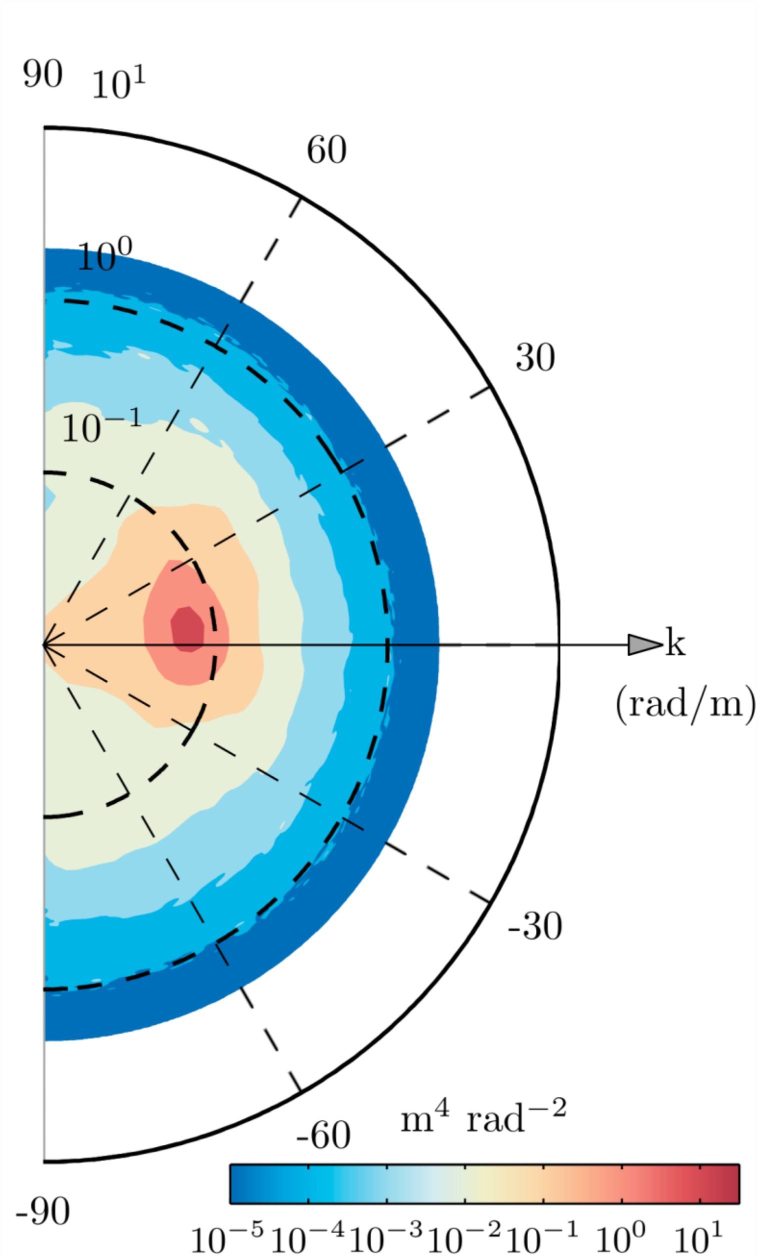


Fig. 4. Sample directional wavenumber spectrum collected at the edge of the Loop Current on Oct. 30th, 2011. The spectrum is rotated into the wind direction (towards 233° from true north) so the wind direction is now θ=0°.

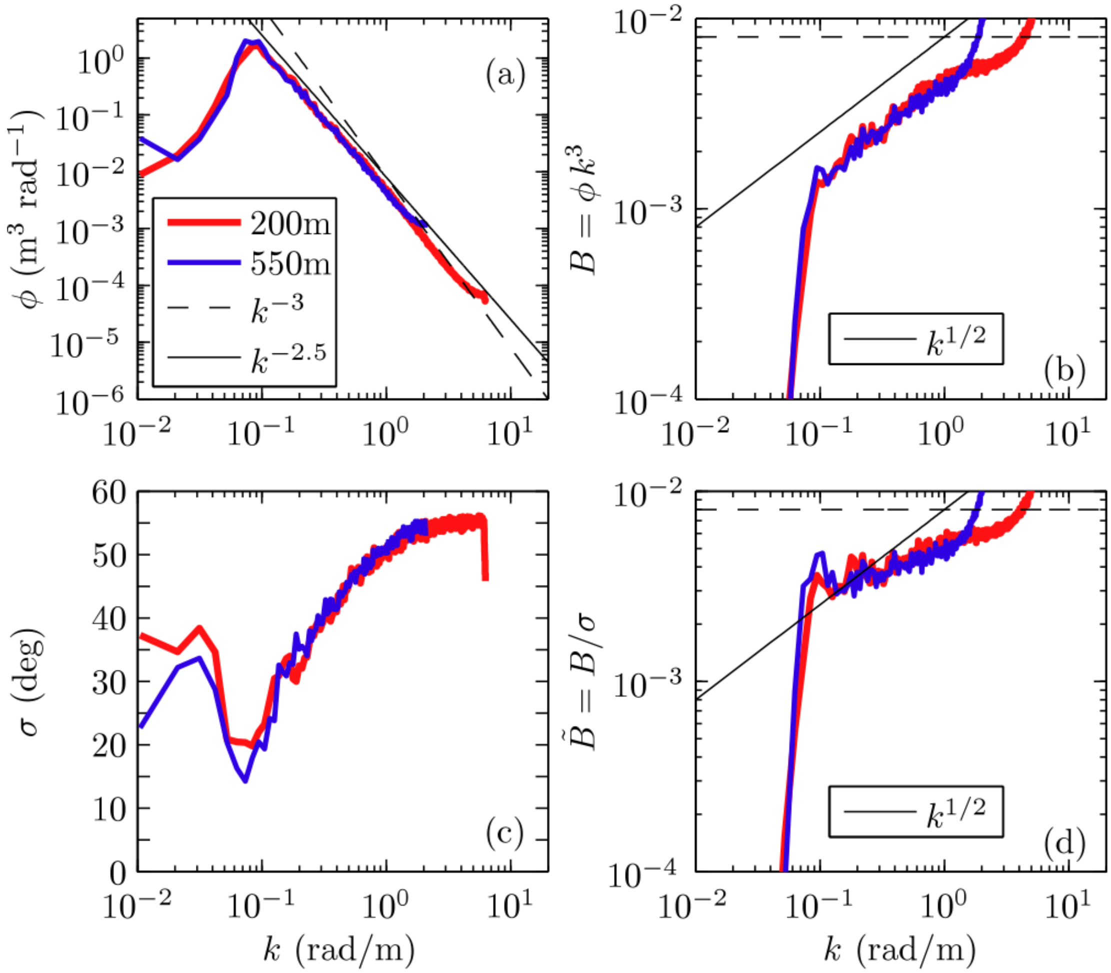


Fig. 5. Sample wavenumber spectra and their moments from lidar measurements collected at the edge of the Loop Current. (a) Azimuth integrated spectra $\phi(k)$, (b) spectral saturation $B(k)= \phi(k)\, k^3$, (c) directional spreading $\sigma(k)$, and (d) normalized saturation $\tilde{B}=B/\sigma$. The red and blue lines correspond to measurements collected at 200 and 550 m, respectively, above MSL. Note that for convenience σ in (c) is in degrees, while in (d) it is in radians. In panel (a) the black solid and dashed lines are reference power-laws of $k^{-2.5}$ and k^{-3}, respectively.

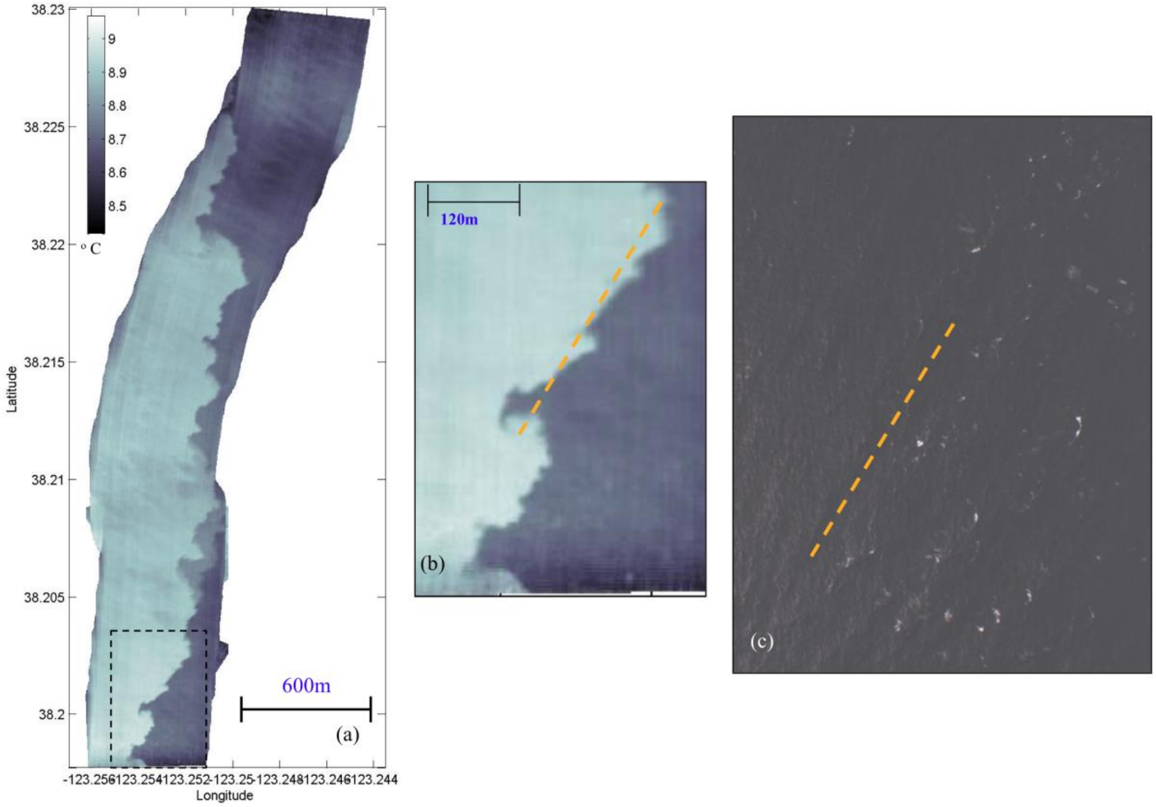


Fig. 6. (a) Georeferenced sea surface temperature map from airborne IR imagery during HiRes on June 17, 2010, with horizontal resolution of 2 m. The black dashed box indicates the corresponding zoomed-in panel (b). Corresponding downward looking airborne visible image is shown in panel (c). The yellow dashed line shows the approximate location of the submesoscale front in (a) and (b).

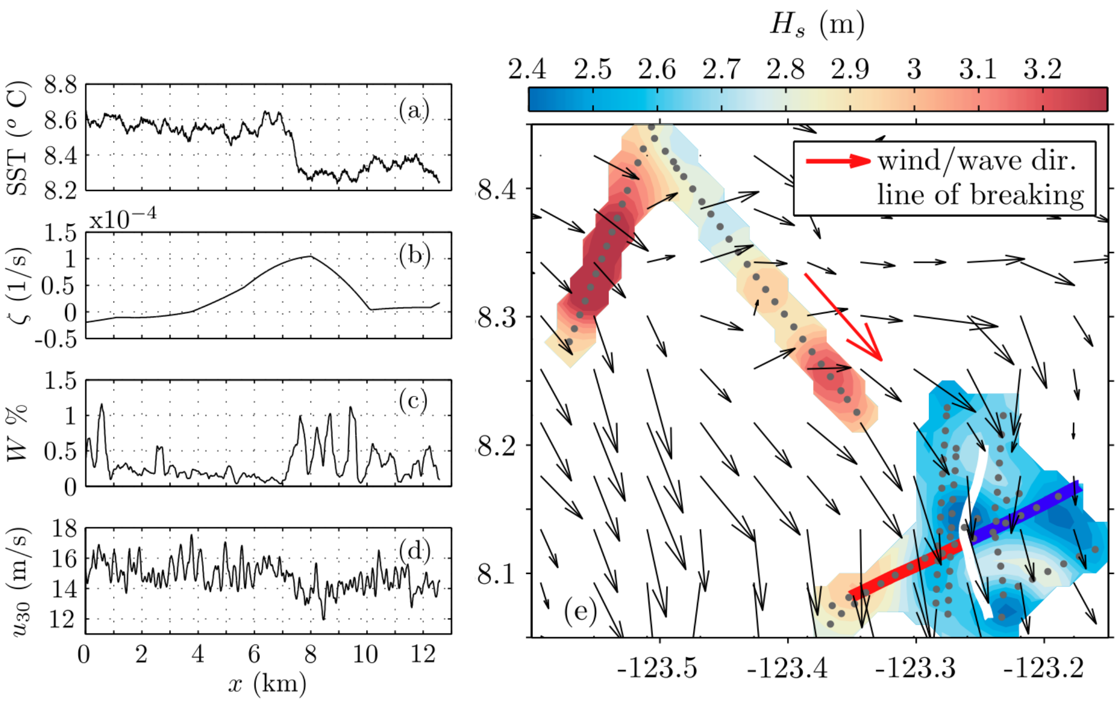
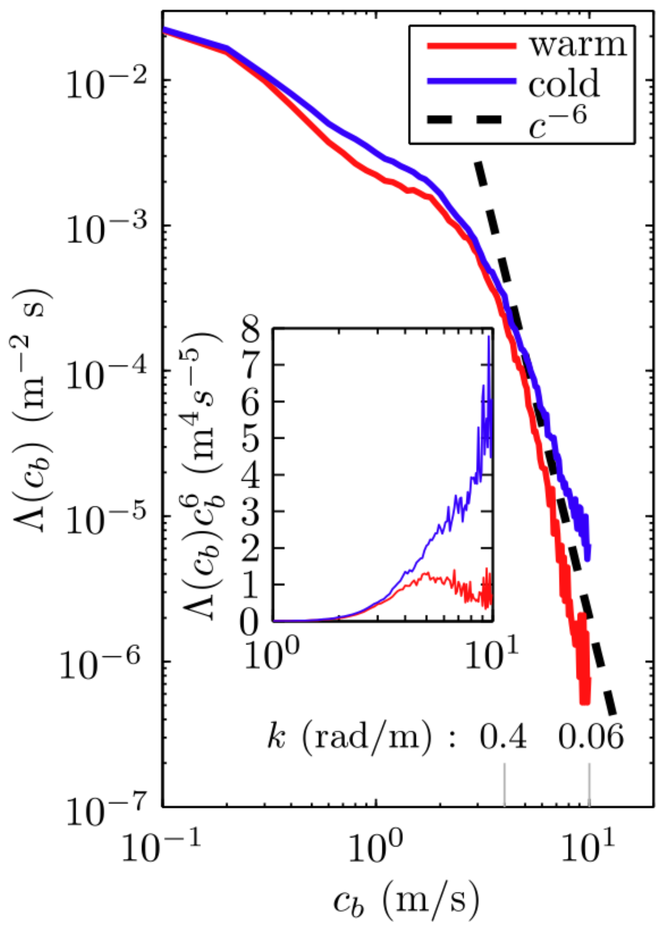


Fig. 7. Spatial cross-front distributions of SST (a), surface vorticity $\zeta = v_x - u_y$ from the HF radar surface currents at 2 km resolution (b); whitecap coverage W (c), and wind speed at 30 m above MSL measured by the Twin Otter aircraft. The position x=0 corresponds to the SW end of the sampling track in panel (e) shown with red and blue lines. (e) Objectively mapped significant wave height from ATM measurements, with the thick white line showing the location of the area with enhanced wave breaking (Fig. 1). The red arrow shows the mean wind and wave direction. The peak wavelength is 112 ± 14 m. The black arrows show current vectors, decimated by a factor of two, from the HF radar data. The gray dots show the ATM sampling locations. The thick red and blue lines indicate sampling tracks with visible imagery used to process the $\Lambda(c_b)$ distributions shown in Figure 8 corresponding to the warm and cold SST areas, respectively.



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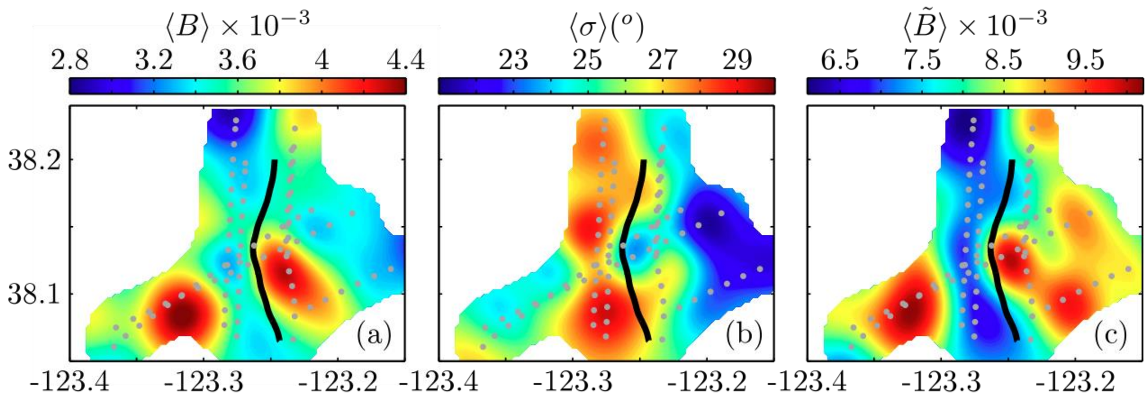

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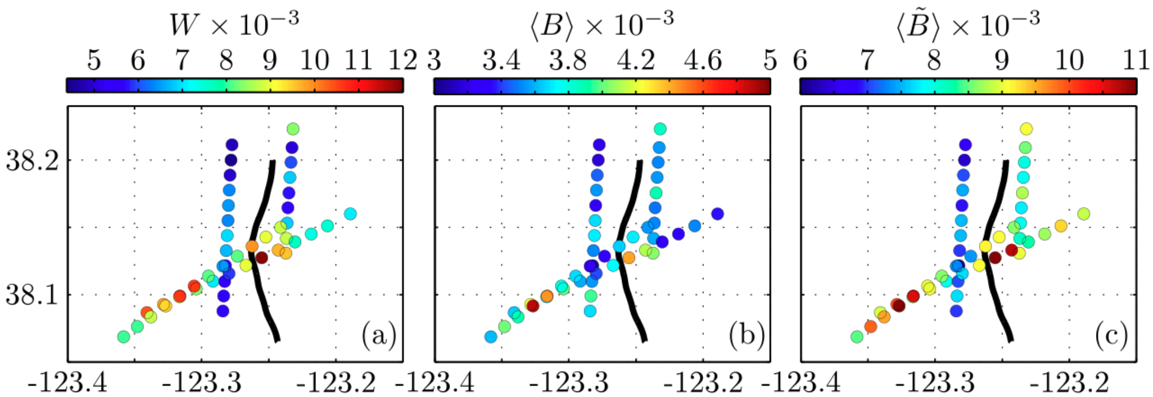

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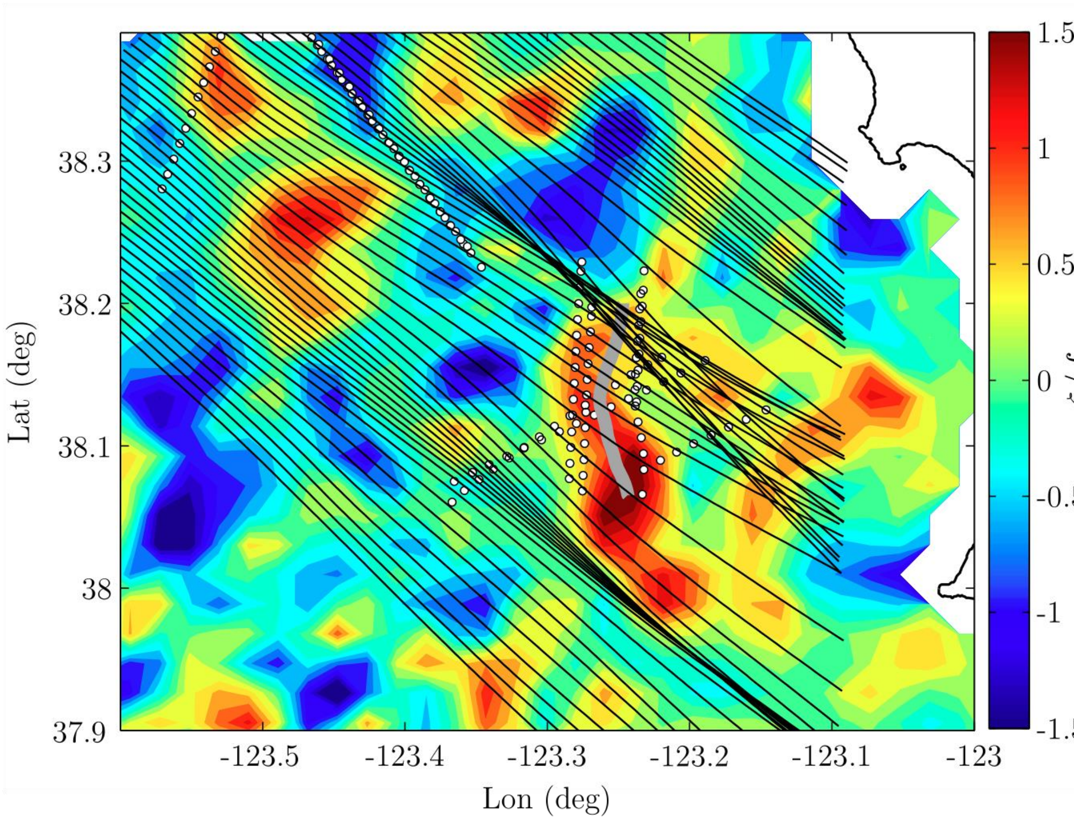

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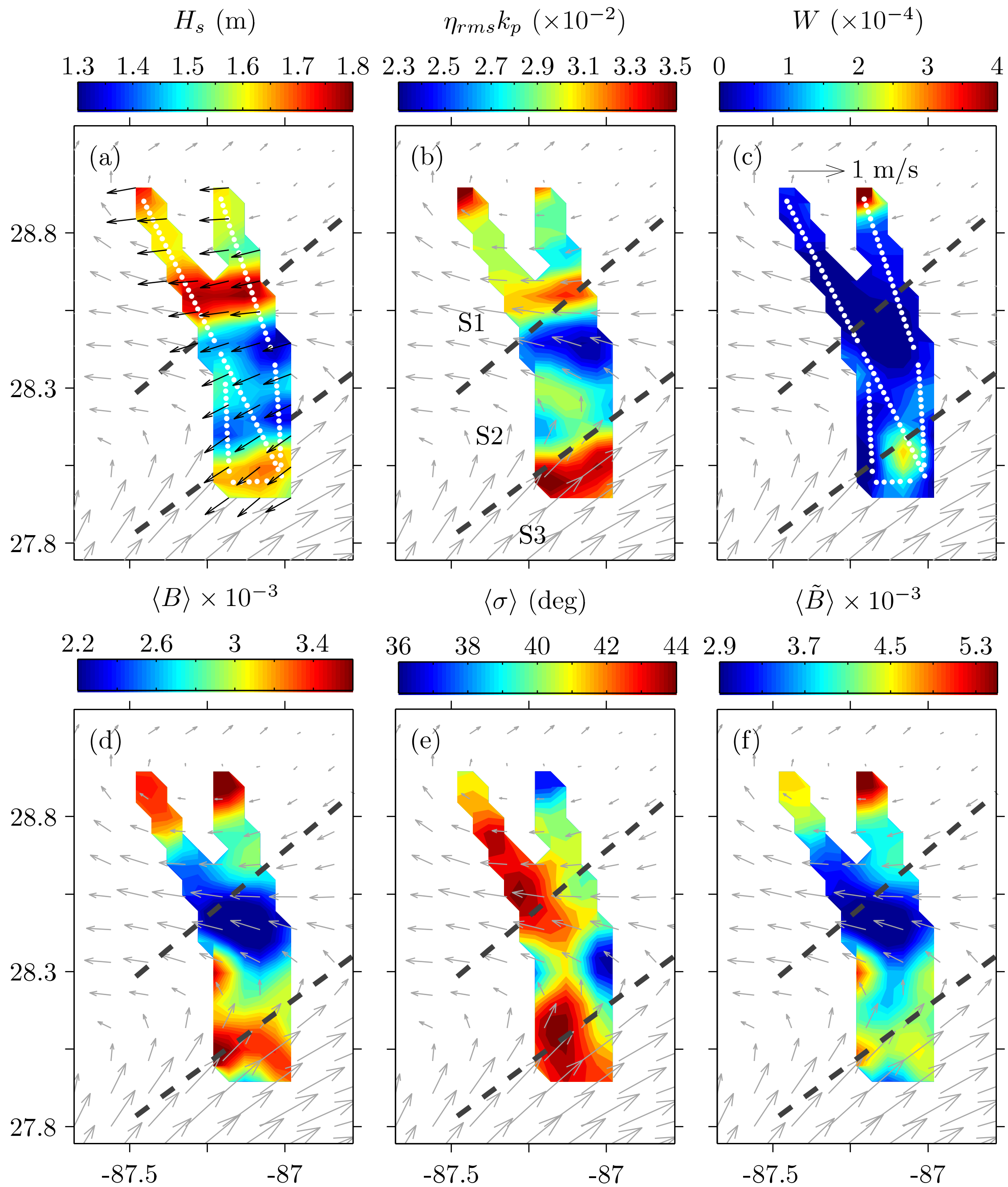

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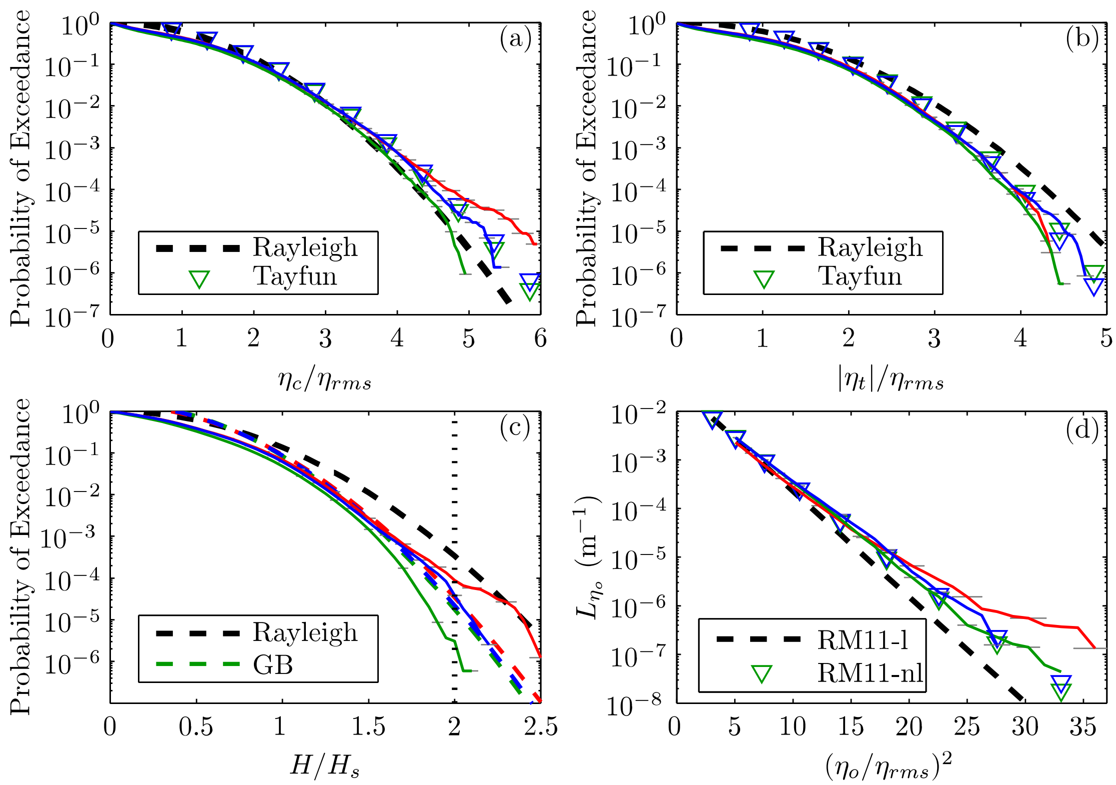

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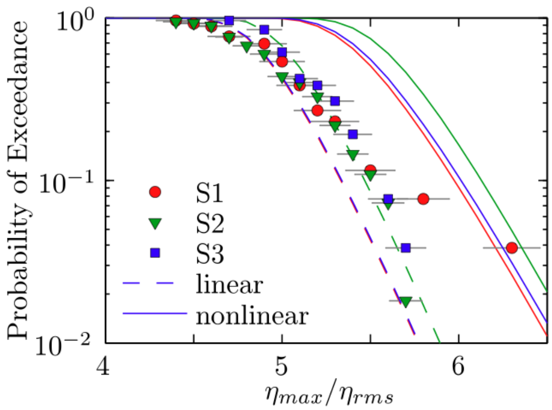

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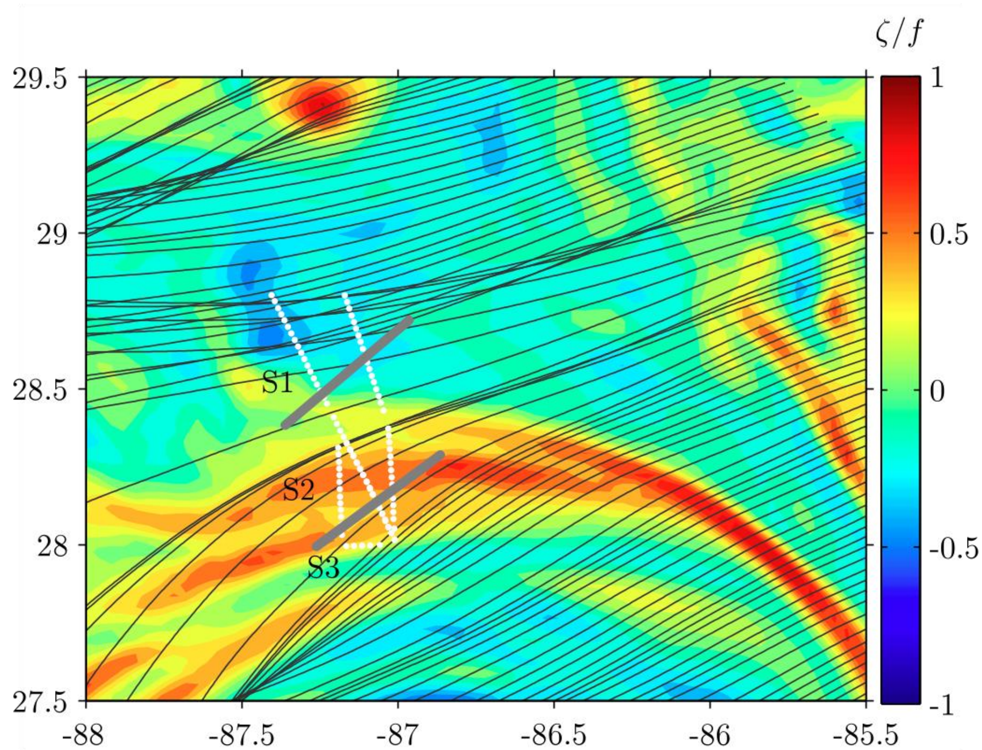

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