Observing Surface Wave Directional Spectra under Typhoon Megi (2010) Using Subsurface EM-APEX Floats

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ABSTRACT: EM-APEX floats as autonomous vehicles have been used for profiling temperature, salinity, and current velocity for more than a decade. In the traditional method for processing horizontal current velocity from float measurements, signals of surface wave motion are removed as residuals. Here, a new data processing method is proposed for deriving the horizontal velocity of surface waves at the floats. Combined with the vertical acceleration measurements of waves, surface wave directional spectra $E(f, \theta)$ can be computed. This method is applied to the float measurements on the right of Typhoon Megi's 2010 track. At 0.6 days before the passage of Megi's eye to the floats, the fast-propagating swell may affect wind waves forced by the local storm wind. When the storm moves closer to the floats, the increasing wind speed and decreasing angle between wind and dominant wave direction may enhance the wind forcing and form a monomodal spectrum $E(f)$. The peak frequency $f_p \sim 0.08$ Hz and significant wave height > 10 m are found near the eyewall. After the passage of the eye to the floats, f_p increases to > 0.1 Hz. Although $E(f)$ still has a single spectral peak at the rear-right quadrant of Megi, $E(f, \theta)$ at frequencies from 0.08 to 0.12 Hz has waves propagating in three different directions as a tri-modal spectrum, partially due to the swell propagating from the rear-left quadrant. Enhancing the capability of EM-APEX floats to observe wave spectra is critical for exploring the roles of surface waves in the upper ocean dynamics in the future.

KEYWORDS: Atmosphere-ocean interaction; Extreme events; Sea state; Tropical cyclones; In situ oceanic observations

1. Introduction

Surface gravity waves at the air–sea interface have critical influences on the momentum transfer from the wind to ocean current, via either the modification on ocean surface roughness length (Smith et al. 1992) or wave breaking (Melville and Rapp 1985). The Stokes drift velocity of surface waves can lead to Langmuir circulation (Craik and Leibovich 1976) and induce turbulent mixing for enhancing the entrainment of cold water from the seasonal thermocline, and thereby the cooling of sea surface temperature. Therefore, recent model studies have extensively coupled the surface wave field into the air–sea interaction processes for forecasting extreme weather systems, such as tropical cyclones (e.g., Moon et al. 2004; Chen et al. 2013). Because it is challenging to measure surface waves in extreme environments under tropical cyclones, simulating storm-induced surface waves relies on the parameterizations applicable in weak wind regimes mostly. Here, a new method is presented for measuring surface wave directional spectra using electromagnetic autonomous profiling explorer (EM-APEX) float measurements. Direct observations on surface waves are

crucial for exploring their roles in atmosphere and ocean boundary layers.

Most previous studies use buoy-mounted accelerometers to measure surface wave energy spectra $E(f)$ (e.g., Graber et al. 2000; Dietrich et al. 2011; Drennan et al. 2014; Collins et al. 2014). Sensors are often mounted along the cable between the buoy and mooring to measure temperature, salinity, and velocity. Unfortunately, the mooring cable is vulnerable to strong vibration induced by breaking waves and turbulence (e.g., under tropical cyclones; Potter et al. 2015). Autonomous surface vehicles (ASV) drifting at the sea surface, such as Wave Gliders (Lenain and Melville 2014) or drifters (Thomson et al. 2018), have also been used for measuring surface waves. Although ASV can be used for studying surface wave spectra under extreme environments (Lenain and Melville 2014), sensors mounted at their base may measure only temperature and salinity near the sea surface. Recently, subsurface Lagrangian floats are used for measuring the $E(f)$ for the first time (D'Asaro 2015). Similar to the ASV, these floats drifting with the seawater motion can profile the upper ocean under tropical cyclones reliably (e.g., Sanford et al. 2011). In other words, subsurface floats can be used for measuring the surface waves and upper ocean structure simultaneously.

Unlike the wave height and wavelength directly affecting the roughness length to the surface wind stress, the propagation direction of surface waves may alter the wind forcing to the surface wave field, and thereby the ocean current (Chen et al. 2013; Hsu et al. 2019). Surface wave directional spectra $E(f, \theta)$ are often estimated (e.g., Kuik et al. 1988; Thomson et al. 2018) to explore the propagation direction of surface waves by computing the phasing between the vertical and horizontal motion of surface waves in the cross-spectra. Because the

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FIG. 1. (left) Structure of EM-APEX float and (right) the schematic illustration on two pairs of electrodes E_1 and E_2 viewed from the top of the float. The electric current induced by the seawater motion is measured by the electrodes as the voltage $(-\mathbf{J}/\sigma = -J_x/\sigma\hat{i} - J_y/\sigma\hat{j})$. The h_x and h_y are the orthogonal axes for the magnetometer measurements. The angle between the h_x and the zonal direction $\hat{i}$ is $\Omega t + \phi_0$, where Ω is the rotation rate of the electrodes.

vertical acceleration of surface waves can be measured by the accelerometers mounted on the free-drifting vehicles accurately, observing the horizontal motion of surface waves is the major challenge for those interested in estimating the $E(f, \theta)$ (Thomson et al. 2018).

Subsurface EM-APEX floats are first designed to measure the ocean current, by capturing the electric current induced by the ocean current in the Earth magnetic field (Sanford et al. 1978, 2005). Though the floats also capture the horizontal electric current induced by the motion of surface waves, the signals are often treated as residuals during the data processing. Hsu et al. (2018) nonlinearly fit the profiles of residuals for estimating the bulk properties of surface waves under a tropical cyclone. It reveals the possible method for measuring the horizontal velocity of surface waves at the EM-APEX floats directly. Furthermore, the accelerometers mounted on the EM-APEX floats can measure the vertical acceleration of surface waves. Using both the horizontal velocity and vertical acceleration of surface waves, one can derive $E(f, \theta)$ from EM-APEX float measurements.

Therefore, the key question is how to develop a reliable method for deriving the horizontal velocity of surface waves from float measurements. This study will propose a new data processing method for decoding and interpreting the signals of surface waves first. The processed measurements will be used for estimating $E(f, \theta)$. Note that EM-APEX floats have been used for measuring temperature, salinity, current velocity (Sanford et al. 2005), and turbulent mixing (Lien et al. 2016) previously. With this new method, EM-APEX floats will be helpful in future studies for exploring the roles of surface waves in the upper ocean. Seven EM-APEX floats were deployed under Typhoon Megi during the Impact of Typhoons on the Ocean in the Pacific (ITOP) experiment in 2010 (D'Asaro et al. 2014b; Hsu et al. 2017). The $E(f, \theta)$ under Typhoon Megi will be estimated using float measurements, and then compared with simulated surface waves in the

surface wave model WAVEWATCH III (WAVEWATCH III Development Group 2016). Unfortunately, there is no direct measurement of surface wave directional spectra taken by other platforms for intercomparison with our float observations.

The structure of EM-APEX floats will be reviewed in section 2. Section 3 will describe the method for estimating the horizontal velocity of surface waves, and thereby the surface wave directional spectra. Float measurements under Megi will be described in section 4. Observations and model results of surface waves will then be discussed in section 5. Section 6 will explore the uncertainties of surface wave estimates using simulated float measurements.

2. EM-APEX floats

Each EM-APEX float is equipped with two pairs of orthogonal Ag–AgCl electrodes E_1 and E_2 , which measure the voltage induced by the seawater motion (Fig. 1; Sanford et al. 2005). An electronic board including both the accelerometer and magnetometer is mounted near the bottom of the floats. The vertical acceleration measured by the accelerometer is sampled by a 10-bit analog to digital (Carlson et al. 2006), ranging within $\pm 2g$ (g is the gravity). The voltage, vertical acceleration, and magnetic field data are output in 1 Hz. The typical vertical profiling speed of the float W is $\sim 0.11 \text{ m s}^{-1}$, slightly faster in descending than ascending (Hsu et al. 2018). As floats profile vertically, their relative vertical motion to the surrounding seawater will force the sensors to rotate around its axis (online supplemental material section A), via an array of slanted blades mounted on the floats (Fig. 1; Hsu et al. 2018). Same as the electric current measured by the orthogonal electrodes (Hsu et al. 2018), magnetic field measurements are projected to the orthogonal axes as two horizontal components h_x and h_y . Because the h_x and h_y are mainly constituted by the horizontal Earth magnetic field F_y , the orientation of EM sensors can be estimated using the two-argument arctangent

FIG. 2. Flowchart for estimating the ensemble mean of surface wave directional spectra $E(f, \theta)$ using the raw measurements taken by EM-APEX floats. Details for processing the float measurements and computing surface wave spectra are described in section 3.

function, i.e., $(\Omega t + \phi_0) = \text{atan2}(h_x, h_y)$, where Ω is the rotational rate of the EM sensors, t is time, and ϕ_0 is a reference phase of h_x at $t = 0$.

3. Method for estimating surface waves using EM-APEX float measurements

This section will introduce the method for deriving surface wave directional spectra at EM-APEX floats, which involves the processing of raw measurements and estimation of surface waves (Fig. 2). The results of wave spectra can then be used for computing the bulk properties of surface waves. Below, how to process the horizontal velocity and vertical acceleration of surface waves will be described first. The processed measurements are then demeaned, detrended (linear), tapered (Hanning), and rescaled to conserve the variance (Thomson et al. 2018). Once the steps for processing the wave measurements are finished, the ensemble

mean of the surface wave energy spectra $\bar{E}(f)$ and normalized directional moments (Thomson et al. 2018) can be computed, respectively. Note that the maximum entropy method (MEM; Lygre and Krogstad 1986) has been used for estimating the $E(f, \theta)$ extensively (Benoit 1992; Young 1999; Collins 2014; Thomson et al. 2018), assuming waves within the same frequency band propagating toward the same direction. A surface wave directional spectrum $E(f, \theta)$ can then be reconstructed based on the MEM method. An example for deriving the $\bar{E}(f, \theta)$ from the raw measurements will be demonstrated in section 3c.

a. Float measurements of surface waves

1) HORIZONTAL VELOCITY OF A SURFACE WAVE

Hsu et al. (2018) describe the electric current $\mathbf{J}$ induced by a single surface wave in the deep water based on the motional induction theory (Table 1; Sanford 1971), i.e.,

TABLE 1. Notations used in this study.

Notation	Definition
$E(f)$	Surface wave energy spectrum
$E(f, \theta)$	Surface wave directional spectrum
Ω	Rotation rate of EM sensors
W	Vertical profiling speed of the floats
F_y and F_z	Horizontal and vertical Earth magnetic fields
$\mathbf{J}$	Electric current induced by surface waves
ω	Wave angular frequency
θ	Direction for waves propagating toward
T	Length of the windows for computing spectrograms
ϵ_i	Residuals in the harmonic fit, mostly constituted by the wave-induced voltage measurements on the electrodes E_i ($i = 1, 2$)
χ_i	χ_i as a constant is the rotational demodulation effect to the float-measured $\mathbf{J}$
β	Geomagnetic field's inclination effect
$\psi_0, \psi, \psi_\omega$, and ψ_Ω	Phase of h_x to the zonal direction at time $t = 0$ (Fig. 1), phase of surface waves in time, phase of wave-induced $\mathbf{J}$, and relative phase of wave propagation direction to the rotating E_1^+ , respectively
f_c	Cutoff frequency of the high-pass filter
u, v , and a	Estimates of zonal velocity, meridional velocity, and vertical acceleration of surface waves, respectively
E'_H and E'_Z	Apparent spectrograms computed using the float measurements of voltage and vertical acceleration, respectively
E_H and E_Z	Adjusted spectrograms derived by correcting E'_H and E'_Z , respectively
θ_1 and $\Delta\theta_1$	Propagation direction and spread of dominant waves at every frequency
R	Check factor, the ratio of $\overline{E_H}$ to $\overline{E_Z}$
-	Ensemble mean of the variables
$C_{\max}$	Criteria for restricting the correction on the depth-decaying effect
Z	Median depth of measurements in the windows (or spectrograms), positive downward
f_p	Peak frequency as the frequency at the peak of $E(f)$
H_s	Significant wave height
θ_p and $\Delta\theta_p$	Propagation direction and spread of dominant waves at f_p
σ_u^2 and $\Delta\sigma_u^2$	σ_u^2 is the variance of residuals in the harmonic fit, which is in the unit of velocity variance ($\text{m}^2 \text{s}^{-2}$); $\Delta\sigma_u^2$ is the absolute difference of σ_u^2 between E_1 and E_2
X	Zonal difference of positions from the floats to Megi's eye; $X < 0$ when Megi's eye has not passed the floats

$$\frac{\mathbf{J}}{\sigma} = \frac{J_x}{\sigma} \hat{\mathbf{i}} + \frac{J_y}{\sigma} \hat{\mathbf{j}} = \sqrt{2} \sigma_{u0} F_z \sqrt{1 + \beta^2} \sin \psi_\omega (\sin \theta \hat{\mathbf{i}} - \cos \theta \hat{\mathbf{j}}) e^{kz}, \quad (1)$$

where σ is the electrical conductivity of seawater, σ_{u0}^2 the variance of horizontal velocity of a surface wave, $k = (k_x^2 + k_y^2)^{1/2}$ the wavenumber magnitude, $\theta = \tan^{-1}(k_y/k_x)$ the wave propagation direction, $\psi = k_x x + k_y y - \omega t + \psi_0$ the phase of a surface wave, $\psi_\omega = \psi + \tan^{-1}\beta$, ω the wave frequency, $\beta = (F_y/F_z) \sin \theta$, $\mathbf{F} = F_y \hat{\mathbf{j}} + F_z \hat{\mathbf{k}}$ the Earth magnetic field, ψ_0 the initial phase of surface wave, $\hat{\mathbf{i}}$, $\hat{\mathbf{j}}$, and $\hat{\mathbf{k}}$ the unit vectors in the x , y , and z directions, respectively, and z is positive upward. The amplitude of electric current is affected by not only the velocity variance of surface waves, but also the geomagnetic field's inclination effect β . The phase of the wave-induced electric current ψ_ω is slightly affected by the β effect, not the same as the phase of vertical acceleration of waves (Hsu et al. 2018). The F_y and F_z are 38 037 and $-16\,052$ nT under Typhoon Megi, respectively, according to the NOAA National Centers for Environmental Information (NCEI) (Thébault et al. 2015).

Two pairs of electrodes E_i ($i = 1$ or 2) on the EM-APEX floats are used to measure the electric current induced by the seawater motion $\mathbf{J}$ (Fig. 1). The voltage measurements are harmonic-fitted in the moving windows with the length of T ($T = 120$ s with 110 s overlapped), assuming the voltage due to the offset, trend, and low-frequency ocean current constant.

Setting $T = 120$ s makes the length of windows longer than that used in the previous studies (e.g., Sanford et al. 2005), in order to decrease the frequency increment in the wave energy spectrum (~ 0.0083 Hz; section 5). The ϵ_i (unit: V) as the residuals of the fitting is mainly due to the high-frequency surface waves and instrumental noise (Hsu et al. 2018), i.e.,

$$\epsilon_i/L = -(1 + C_1) \left(\frac{J_x}{\sigma} \cos \theta_\Omega + \frac{J_y}{\sigma} \sin \theta_\Omega - \chi_i + \delta \right), \quad (2)$$

where L is the separation between the electrode pairs (~ 0.22 m), $C_1 \sim 0.5$ the head factor which enhances the electric current surrounding the floats, $\theta_\Omega = \Omega t + \phi_0 - (\pi/2)(i - 1) + \alpha_1$ the orientation of E_i^+ to the magnetic east, α_1 the angle clockwise from E_1^+ to h_x , χ_i the average of $A \cos(\psi_\omega - \psi_\Omega)$ over the period of T , $\psi_\Omega = \theta - \theta_\Omega$, $A = (1/2)\sqrt{2} \sigma_{u0} F_z \sqrt{1 + \beta^2} e^{kz}$, z the depth of wave measurements, and δ the instrumental noise, assuming the noise on E_1 and E_2 is the same. The χ_i equals zero, unless $\omega \approx \Omega$, termed the rotational demodulation effect in Hsu et al. (2018). The ϵ_i can be regarded as the projection of $\mathbf{J}$ on the orthogonal electrodes, i.e., in the rotating frame. After substituting Eq. (1) into Eq. (2), we can rewrite the ϵ_i (Hsu et al. 2018) as

$$\epsilon_i/L = -(1 + C_1) \{ A [\cos(\psi_\omega - \psi_\Omega) - \cos(\psi_\omega + \psi_\Omega)] - \chi_i \} + \delta. \quad (3)$$

Because signals of surface waves in float measurements are modulated by the rotating electrodes, the surface wave components of ϵ_i include three parts: low-frequency (the difference between wave frequency and rotation frequency) and high-frequency portion (the sum of wave frequency and rotation frequency) of wave-induced voltage measurements, and the χ_i .

A second-order Butterworth high-pass filter is applied to ϵ_i [Eq. (3)] for deriving the filtered voltage measurements ϵ'_i , by using the cutoff frequency $f_c = f_0 + \Omega/2\pi$ Hz ($f_0 = 0.1$ Hz). The main purpose of high-pass filtering is to exclude the constant χ_i (rotational demodulation effect) and the low-frequency portion of ϵ'_i (due to the frequency difference). It can also filter out the non-wave-induced voltage that has not been removed by the harmonic fit. Assuming both the χ_i and low-frequency portion of ϵ'_i removed reliably, Eq. (3) can be rewritten as

$$\epsilon'_i/L \approx (1 + C_1)[A \cos(\psi_\omega + \psi_\Omega) + \delta]. \quad (4)$$

Filtered measurements at two edges of the windows with the length of $1/f_c$ are excluded due to the transient effect. Performing a rotation matrix

$$\mathbf{M} = \begin{bmatrix} \cos\theta_\Omega & -\sin\theta_\Omega \\ \sin\theta_\Omega & \cos\theta_\Omega \end{bmatrix}$$

onto ϵ'_i , the filtered measurements on the Cartesian coordinates $H' = H'_x \hat{\mathbf{i}} + H'_y \hat{\mathbf{j}}$ can be computed as

$$\begin{bmatrix} H'_x/L \\ H'_y/L \end{bmatrix} = \mathbf{M} \begin{bmatrix} \epsilon'_2/L \\ \epsilon'_1/L \end{bmatrix} = (1 + C_1) \begin{bmatrix} -A \sin(\psi_\omega + \theta) + \delta \\ A \cos(\psi_\omega + \theta) + \delta \end{bmatrix}. \quad (5)$$

The variance of H' equals the half of the variance of ϵ_i [Eq. (2)], so the energy spectrum can be computed by using H' . Unfortunately, because the phase of ϵ_i will be modulated at the same time (the same magnitude between H'_x and H'_y), the H' cannot be used for estimating the propagation axis of surface waves.

To derive the measurements for estimating wave propagation direction, the rotation matrix $\mathbf{M}$ is applied to the ϵ_i [Eq. (2)] without using the high-pass filter. The unfiltered voltage measurements $H = H_x \hat{\mathbf{i}} + H_y \hat{\mathbf{j}}$ on the Cartesian coordinates can be computed as

$$\begin{bmatrix} H_x/L \\ H_y/L \end{bmatrix} = \mathbf{M} \begin{bmatrix} \epsilon_2/L \\ \epsilon_1/L \end{bmatrix} = -(1 + C_1) \begin{bmatrix} -\frac{J_y}{\sigma} + \alpha A \sin(\psi_p - \theta_\Omega) + \delta \\ \frac{J_x}{\sigma} + \alpha A \cos(\psi_p - \theta_\Omega) + \delta \end{bmatrix}, \quad (6)$$

where

$$\alpha \approx \begin{cases} 1 & \text{if } |\omega - \Omega| < 2\pi/T \\ 0 & \text{if } |\omega - \Omega| \geq 2\pi/T \end{cases},$$

$\psi_p = \tan^{-1}[\sin(\psi_\omega - \theta + \theta_\Omega)/\cos(\psi_\omega - \theta + \theta_\Omega)]$, and $\langle \cdot \rangle$ the mean over the period of T . Using the expression of H in Eq. (6), the surface wave velocity $\mathbf{V} = u \hat{\mathbf{i}} + v \hat{\mathbf{j}}$ can be estimated as

$$\begin{cases} u = -\frac{H_x}{F_z(1 + C_1)L} \approx \sqrt{2}\sigma_{u0}\sqrt{1 + \beta^2}e^{kz} \left[\sin\psi_\omega \cos\theta - \frac{\alpha}{2} \sin(\psi_p - \theta_\Omega) + \delta_u \right] \\ v = -\frac{H_y}{F_z(1 + C_1)L} \approx \sqrt{2}\sigma_{u0}\sqrt{1 + \beta^2}e^{kz} \left[\sin\psi_\omega \sin\theta + \frac{\alpha}{2} \cos(\psi_p - \theta_\Omega) + \delta_v \right] \end{cases}. \quad (7)$$

The instrumental noise of velocity (zonal δ_u and meridional δ_v) is $\sim 0.008\text{--}0.015 \text{ m s}^{-1}$ (Hsu et al. 2017). Here, we derive the analytical solutions of horizontal velocity assuming a single surface wave, in order to better present how the rotation of EM sensors can modulate the measured surface wave whose frequency is similar with the $\Omega/(2\pi)$. Because the float measurements include multiple waves, the steps for correcting the estimated spectra will be described in the appendix.

2) VERTICAL ACCELERATION OF SURFACE WAVES

The vertical acceleration of surface waves measured by an accelerometer is often used for estimating surface wave energy spectra (e.g., Collins et al. 2014). Though the pressure perturbation caused by surface waves is used for measuring the vertical motion on subsurface floats in D'Asaro (2015), the accelerometer is also recommended as an alternative option for wave measurements. The vertical acceleration measurements a (unit: m s^{-2}) in each profile are parsed into the moving windows of 120 s ($=T$) with 110 s overlapped, and then processed using a

classic high-pass filter (assuming $f_c = 0.02$ Hz; Thomson et al. 2018) as

$$a(t) = \gamma a(t - \Delta t) + \gamma[a(t) - a(t - \Delta t)]; \quad \gamma = \frac{(2\pi f_c)^{-1}}{(2\pi f_c)^{-1} + \Delta t}, \quad (8)$$

where Δt is the time interval in the vertical acceleration measurements ($=1$ s). If we choose to use the Butterworth high-pass filter for processing the vertical acceleration measurements, the same as that performed onto the voltage measurements [Eq. (4)], more than 80% of data points need to be excluded in each window to avoid the transient effect.

b. Surface wave directional spectrum

1) SURFACE WAVE ENERGY SPECTRUM

We compute the apparent spectrograms E'_H and E'_Z using the processed measurements of voltage [H' in Eq. (5); unit: volt] and vertical acceleration a via the fast Fourier transform (Thomson et al. 2018), respectively, i.e.,

$$\begin{cases} E'_H = \zeta \frac{S'_{HH}}{\omega^2 \left\{ 1 + \left[\frac{F_y}{F_z} \sin \bar{\theta}_1(f) \right]^2 \right\} F_z^2 (1 + C_1)^2 L^2}; & \begin{cases} \zeta = 2 & \text{at } f < f_c \\ \zeta = 1 & \text{at } f \geq f_c \end{cases} \\ E'_Z = \frac{S'_{aa}}{\omega^4} \end{cases}, \quad (9)$$

where S'_{HH} and S'_{aa} are the autospectra computed using the H' and a , respectively, and $\bar{\theta}_1(f)$ the ensemble mean of dominant wave direction, which is estimated by using Eq. (13). Note that we use the Butterworth high-pass filter in the processing of the voltage measurements [Eq. (4)] to exclude the α effect [i.e., χ_i in Eq. (3)], which will remove the half of variance of low-frequency waves at the same time (supplemental material, section C). The factor ζ therefore equals 2 at $f < f_c$. Because the variance of high-frequency waves at $f \geq f_c$ should not be affected significantly, the $\zeta = 1$ at $f \geq f_c$.

Because surface waves decay exponentially in depth at the deep water, termed the depth-decaying effect, the E'_H and E'_Z estimated using float measurements at the deeper layers are underestimated and requires appropriate corrections. The adjusted spectrograms E_H and E_Z are derived after correcting the E'_H and E'_Z , by following the steps described in D'Asaro (2015) (appendix). One of the most important corrections is to amplify the spectral level due to the depth-decaying effect. A parameter $C_{\max}$ is used to exclude the unrealistically amplified spectral level at the frequency $> \sqrt{g \log(C_{\max}) / [(2\pi)^2 Z]}$ Hz, where Z is the median depth of measurements of the spectrograms (positive downward). The spectrograms at the deeper layers will have fewer estimates at the high-frequency bands. This study uses $C_{\max} = 10$ (section 6), the same as that in D'Asaro (2015), because larger $C_{\max}$ may fail to remove biases at the high-frequency bands.

All adjusted spectrograms are then averaged to derive the ensemble mean of wave energy spectra $\bar{E}(f)$. We randomly select and average 80% of spectrograms to compute the spectral average 100 times, similar with the bootstrap method used in Hsu et al. (2018). The mean and standard deviation of $\bar{E}(f)$ at each frequency band are computed by using these 100 realizations of spectral averages. The $\bar{E}(f_s)$ is excluded if there are less than four spectrograms with available estimates at f_s . The significant wave height H_s and peak frequency f_p . The value of H_s can be computed using the $\bar{E}(f)$ (Young 1999) as

$$H_s \approx 4 \sqrt{\int \bar{E}(f) df} \quad \text{and} \quad f_p = f\{\bar{E}(f) = \max[\bar{E}(f)]\}. \quad (10)$$

Based on the linear wave theory, the variance of horizontal motion of surface waves should equal the variance of vertical wave motion at the deep water (Thomson et al. 2018). A check factor R is defined (Thomson et al. 2015) as

$$R = \frac{\bar{E}_H}{\bar{E}_Z}, \quad (11)$$

where $\bar{E}_H$ and $\bar{E}_Z$ are the ensemble mean of E_H and E_Z , respectively. Estimates of surface wave energy spectra are reliable

when $R = 1$. Since the depth of ocean floor (~ 5000 m) is much larger than the typical wavelength of surface waves under tropical cyclones (~ 100 – 300 m), the R will be used as the major indicator to evaluate the quality of surface wave estimates in this study.

2) PROPAGATION DIRECTION OF DOMINANT WAVES AT EVERY FREQUENCY BAND

The $E(f, \theta)$ is often estimated (Kuik et al. 1988; Herbers et al. 2012; Thomson et al. 2018) by computing the phasing between the vertical and horizontal motion of surface waves in the cross-spectra. In this study, the unfiltered voltage measurements H in each window are used to derive the horizontal velocity of surface waves [u and v in Eq. (7)]. Combined with the processed vertical acceleration measurements a , the normalized directional moments a_1 and b_1 can be computed (Thomson et al. 2018) as

$$\begin{cases} a_1(f) = \frac{\text{Re}(C_{ua})}{\sqrt{(S_{uu} + S_{vv})S_{aa}}} \\ b_1(f) = \frac{\text{Re}(C_{va})}{\sqrt{(S_{uu} + S_{vv})S_{aa}}} \\ a_2(f) = \frac{S_{uu} - S_{vv}}{S_{uu} + S_{vv}} \\ b_2(f) = \frac{2\text{Re}(C_{uv})}{S_{uu} + S_{vv}} \end{cases}, \quad (12)$$

where Re is the real part of the cross-spectra. C_{ua} , C_{va} , and C_{uv} are cospectra (in phase) between u [$\propto \sin \psi \cos \theta$ in Eq. (1)] and a ($\propto \sin \psi$), between v [$\propto \sin \psi \sin \theta$ in Eq. (1)] and a and between u and v , respectively. S_{uu} and S_{vv} are the autospectra of u and v , respectively. The C_{ua} and C_{va} in the a_1 and b_1 are thus proportional to the $\cos(\tan^{-1}\beta) \cos \theta$ and $\cos(\tan^{-1}\beta) \sin \theta$, respectively, where the $\cos(\tan^{-1}\beta)$ is the shifted phase of wave-induced electric current due to the β effect [Eq. (1)]. The normalized directional moments describe the phasing between horizontal and vertical motion of surface waves, independent with the magnitude of wave measurements (Thomson et al. 2018). That is, the estimated magnitude of u and v [Eq. (7)] will not affect the results of normalized directional moments.

The ensemble means of propagation direction $\bar{\theta}_1$ and spread $\Delta \bar{\theta}_1$ of dominant waves at each frequency band are computed using the ensemble mean of a_1 and b_1 (i.e., $\bar{a}_1$ and $\bar{b}_1$), i.e.,

$$\bar{\theta}_1(f) = a \tan 2(\bar{b}_1, \bar{a}_1) \quad \text{and} \quad \Delta \bar{\theta}_1(f) = \sqrt{2 \left(1 - \sqrt{\bar{a}_1^2 + \bar{b}_1^2} \right)}. \quad (13)$$

Voltage Processing

Spectrum Estimates

FIG. 3. Example for demonstrating how to use the raw voltage measurements on E_1 [blue solid line in (a)] and E_2 [red solid line in (a)] to derive the surface wave spectra. (b) The wave-induced voltage is the difference between the raw data and the harmonic fitted results [dashed lines in (a)]. (d) The horizontal velocity of surface waves (u : blue line; v : red line) are derived by rotating (b) to the Cartesian coordinates. After using a high-pass filter onto (b) for deriving (c), the filtered voltage measurements are also (e) rotated to the Cartesian coordinates as H' (H'_x : red line; H'_y : blue line). The normalized directional moments a_1 and b_1 [presented as the wave direction $\theta(f) = \tan^{-1}(b_1/a_1)$ in (f)], E'_H [blue line in (g)] and E'_Z [blue line in (h)] are computed using (d) and (e). The E_H [red line in (g)] and E_Z [red line in (h)] are derived after correcting the E'_H and E'_Z . Using the results of spectrograms in a single profile, the results of (i) $\bar{\theta}(f)$, (j) $\bar{E}(f)$, and (k) $E(f, \theta)$ are presented.

The β effect cannot change the estimates of $\bar{\theta}_1$ by computing the ratio of C_{va} to C_{ua} . The uncertainty due to the α effect can be reduced by computing the ensemble mean (not shown in this study), because the rotation rate of EM sensors Ω varies during profiling. In other words, the results of $\bar{\theta}_1$ can be used for computing the E'_H directly [Eq. (9)]. Then the propagation direction θ_p and spread $\Delta\theta_p$ of surface waves at f_p are estimated by using the $\bar{\theta}_1(f)$ and $\Delta\bar{\theta}_1(f)$, i.e.,

$$\theta_p = \bar{\theta}_1(f_p) \quad \text{and} \quad \Delta\theta_p = \Delta\bar{\theta}_1(f_p). \quad (14)$$

c. Example of deriving spectra from raw voltage measurements

The measurements at the float em4913a (section 4) are used (solid lines in Fig. 3a) to demonstrate the steps for estimating the wave spectra. We harmonic fit the raw voltage data in a 120-s window for estimating the voltage not caused by surface waves first (dashed lines in Fig. 3a). The residuals of the fitted result are derived (Fig. 3b), which should be mostly due to surface waves. The horizontal velocity of surface waves [Eq. (7)] can be computed by performing the rotation matrix $\mathbf{M}$ onto the wave-induced voltage (Fig. 3d). Because of the rotational demodulation effect, a high-pass filter is performed onto the residuals (Fig. 3c). Note that the harmonic fit may not capture the nonlinear change of voltage offset (section 3a). Using the

high-pass filter can further extract the voltage signals due to high-frequency surface waves. The filtered voltage on the Cartesian coordinates H' can then be computed (Fig. 3e).

The measurements of vertical acceleration a and horizontal velocity of surface waves are used for computing the normalized directional moments [Fig. 3f shows $\theta(f) = \tan^{-1}(b_1/a_1)$]. The ensemble-mean $\bar{a}_1$ and $\bar{b}_1$ over the whole profile are used to compute the $\bar{\theta}_1(f)$ (Fig. 3i). The E'_H and E'_Z are computed by using the H' , $\bar{\theta}_1$ and a (blue lines in Figs. 3g,h), respectively. The adjusted spectrograms in the profile are derived by correcting the E'_H and E'_Z (red lines in Figs. 3g,h), following the steps in the appendix. We can then estimate the $\bar{E}(f)$ (Fig. 3j), and thereby the $E(f, \theta)$ via the MEM method (Fig. 3k).

4. Measurements under Typhoon Megi

a. Float measurements

Seven EM-APEX floats were air launched by a C130 aircraft at around 18.7°N, 128.3°E on 16 October 2010, ~1 day before the passage of Typhoon Megi to the floats (D'Asaro et al. 2014b; Hsu et al. 2017). The $|U_{10}|$ at the float positions is not more than 20 m s^{-1} until 1400 UTC 16 October 2010 (Fig. 4). After the passage of the storm, floats were recovered for retrieving the raw data. Measurements taken by two floats em4913a and em3766c, ~37 and 73 km to the right of Megi's

FIG. 4. (a) Wind speed at 10-m height above the sea surface, $|U_{10}|$, under Typhoon Megi at 2030 UTC 16 Oct 2010, and (b) interpolated $|U_{10}|$ at the positions of float em4913a (red lines) and em3766c (blue lines). Details for wind map processing under Megi are discussed in Hsu et al. (2017).

track, respectively, will be used for estimating surface waves. The floats continuously profile in the upper 200 m (Figs. 5a,b). The estimated wave energy spectra $\bar{E}(f)$ extend up to 0.2 Hz by using measurements at each profile (section 5). After 1400 UTC 16 October 2010 (~ 0.4 day before the arrival of

Megi's eye to the floats), the floats profile only at below 30-m depth. Missing measurements in the upper 30 m will restrict our estimates of $\bar{E}(f)$ at the frequency range below 0.13 Hz. Despite missing the high-frequency end, surface wave peak frequencies f_p are captured in our estimates, based on results

FIG. 5. (a),(d) Vertical trajectories of the floats during descending (red dots) and ascending (blue dots), (b),(e) the computed σ_u^2 on E_1 (lines) and E_2 (dots), and (c),(f) the absolute difference of σ_u^2 between E_1 and E_2 (color shading) of two EM-APEX floats (left) em4913a and (right) em3766c. Black lines in (c) and (f) are the estimates of surface mixed layer depth [see Hsu et al. (2017) for more details].

FIG. 6. Surface wave energy spectra $\overline{E(f)}$ and the check factor R estimated using measurements of EM-APEX float (left) em4913a and (right) em3766c. (c),(d) The mean peak frequency f_p (black solid lines) with ± 1 standard deviation (between black solid and dashed lines). The vertical blue dashed lines indicate the time when Megi's eye passed by the floats.

reported by the previous studies (~ 0.08 – 0.1 Hz in Fan et al. 2009) and confirmed in our analysis (Fig. 6).

b. Criteria for quality control

When processing horizontal velocity of seawater, the voltage measurements from floats due to the offset, trend, and low-frequency current are estimated via the harmonic fit, assuming these variables constant within the preset processing time window (Sanford et al. 1978). Residuals of the harmonic fit as the wave-induced voltage measurements ϵ_i are then used for computing the horizontal velocity of surface waves. The variance of velocity residuals σ_u^2 as the variance of $\epsilon_i/[F_z(1+C_1)(1+L)]$ [Eq. (2); Hsu et al. 2018] on the E_1 and E_2 is computed. Most of σ_u^2 profiles near the sea surface decay exponentially in depth (Figs. 5c,d), consistent with that reported by Hsu et al. (2018).

Two criteria are defined to exclude potential factors that may affect the estimation on the horizontal velocity of surface waves. When the float profiles at a slow vertical speed W , the vertical shear may affect the measured h_x and h_y (section 3) by tilting the float, and thereby the orientation of EM sensors θ_Ω [Eq. (2)]. We therefore exclude the spectrograms with $W < 0.06 \text{ m s}^{-1}$ from the analysis. The trend represents the change of voltage affected by the seawater properties and varies between each window (supplemental material, section A). The voltage offset in each window may change nonlinearly, conflicted with the assumptions used in the harmonic fit. Most absolute difference of σ_u^2 between the E_1 and E_2 (i.e., $\Delta\sigma_u^2$) in the surface mixed layer is small, $< 0.05 \text{ m}^2 \text{ s}^{-2}$ (Figs. 5e,f). Some σ_u^2 on the E_1 of em3766c does not decay exponentially in depth, yielding the high $\Delta\sigma_u^2 > 0.2 \text{ m}^2 \text{ s}^{-2}$ below 50-m depth. The sharp change of temperature near the base of the surface mixed layer (black lines in Figs. 5c,f) may bias the data

processing, thereby the results of horizontal wave velocity. About 5% E_H having $\Delta\sigma_u^2 > 0.08$ and $0.49 \text{ m}^2 \text{ s}^{-2}$ in the upper 50 m at em4913a and em3766c, respectively, will be excluded from the following analysis. We also exclude the profiles with no measurements in the upper 50 m.

5. Surface waves under Typhoon Megi

a. Observed surface wave spectra

Surface waves under Typhoon Megi are estimated using the measurements taken by two EM-APEX floats em4913a and em3766c (Fig. 6). Around 1000 UTC 16 October 2010, the f_p is ~ 0.09 Hz at em4913a, slightly lower than that at em3766c. The frequency range of $\overline{E(f)}$ can be up to 0.2 Hz, revealed by measurements available in the upper 30 m. Between 0900 and 1200 UTC 16 October 2010, the $\overline{E(f)}$ at em3766c may have two spectral peaks at 0.08 and 0.12 Hz, and the standard deviation of f_p is > 0.02 Hz. The swell induced by Megi's wind may complicate the surface wave field and result in bimodal spectra, when the eye of Megi is still > 300 km from the float positions.

When Megi moves closer to the floats, the wind speed $|U_{10}|$ increases, and the wind vector rotates clockwise on the right of Megi's track. It continuously forces surface waves and enhances the peak of $\overline{E(f)}$, as the extended fetch effect described in Young (2006). Compared with $\overline{E(f)}$ before 1400 UTC 16 October 2010, the energy of $\overline{E(f)}$ concentrates in a single spectral peak as monomodal spectra (Fig. 6). The f_p is downshifted to 0.08 Hz due to the nonlinear process in the growth of surface waves (Fontaine 2013). The f_p after the passage of Megi's eye is about 0.1 Hz, higher than that at the front-right quadrant of Megi, consistent with the previous studies (Hsu et al. 2018).

FIG. 7. Estimates of surface wave directional spectra $\overline{E(f, \theta)}$ (inset panels) using measurements in each profile at float positions relative to the eye of Typhoon Megi (x axis is the zonal distance, and y axis is the meridional distance). In the inset panels, the radial axis is the frequency (<0.13 Hz), the azimuthal axis is the propagation direction of dominant waves (0° – 360°), and the red arrows are the wind direction. The inner and outer gray circles are the frequencies of 0.06 and 0.1 Hz, respectively. The color shading is the wind speed at 10-m height under Typhoon Megi at 2030 UTC 16 Oct 2010. The tropical cyclone moves westward (to the left, black arrow).

The check factor R around the f_p at most profiles is about 1, except the profiles around 2030 UTC 16 October 2010 (the arrival time of Megi's eye to the floats). The direction of high-frequency wind waves, which should align with the wind vector, may change rapidly near the eyewall of Megi. It may affect the estimates of $\theta_1(f)$ at the high-frequency bands, and then result in the overestimated E_H .

We compute the $E(f, \theta)$ by using the results of $\overline{E(f)}$ and normalized directional moments (Fig. 7). Before the arrival of Megi's eye to the floats (i.e., the zonal distance from the float positions to Megi's eye $X < 0$), the propagation direction of dominant waves is about 180° (westward), and can be up to 80° clockwise from the wind direction. The angle between the wind vector and dominant wave direction decreases with the decreasing $|X|$, consistent with the results reported by the previous model studies (Moon et al. 2004). The alignment between the wind and surface waves near the right-hand side of Megi's eyewall may enhance the growth of long waves propagating into the front-right quadrant leading to bimodal spectra. Surface wave field in the front-right quadrant of tropical cyclones is a combination of wind waves generated by the local wind, and swell generated from the rear-right quadrant, as that reported by Hu and Chen (2011).

At the rear-right quadrant of Megi (Fig. 7), the angle between the wind and dominant surface waves is less than 45° , which is smaller than that at the same distance from Megi's eye in the front-right quadrant. Previous studies conclude that wind waves are more dominant than swell at the rear-right

quadrant of tropical cyclones (Hu and Chen 2011). Interestingly, $E(f, \theta)$ at $X = 58$ km of em4913a exhibits three different propagation directions at the frequency bands from 0.08 to 0.12 Hz. We suspect that surface waves propagating toward 90° may be the swell from the rear-left quadrant of Megi. Because of the similar spectral level at these frequency bands, the $E(f)$ at $X = 58$ km has a single spectral peak that is broader than other spectra at the front of Megi (Fig. 6). That is, surface waves at the rear-right quadrant of Megi may be trimodal in direction instead of frequency, consistent with the results in the previous studies (Wright et al. 2001; Walsh et al. 2002; Black et al. 2007).

b. Observed and simulated bulk properties of surface waves

The bulk properties of surface waves are computed. Because the frequency range of $\overline{E(f)}$ may affect the estimates of H_s , the missing spectral estimates at the high-frequency bands are extrapolated, assuming the wave energy proportional to f^{-4} (the spectrum form in Donelan et al. 1985) or f^{-5} (the JONSWAP spectrum form in Hasselmann et al. 1973). The value of H_s is not affected by the extrapolation significantly. The float-measured surface waves will be discussed and compared with the simulated surface waves in the version 5.16 WAVEWATCH III model (ww3; supplemental material, section B; Hasselmann et al. 1985; Tolman and Chalikov 1996; WAVEWATCH III Development Group 2016), which have been used for studying

FIG. 8. (a) Zonal distance from Megi's eye to float positions X (red lines: observations at em4913a; blue lines: observations at em3766c), (b) significant wave height H_s , (c) peak frequency f_p , (d) dominant wave propagation direction θ_p [solid lines, labels on the left y axis] and its spread at the frequency of dominant waves $\Delta\theta_p$ [dashed lines, labels on the right y axis]. In (b) and (c), the error bars are the standard deviation of surface wave estimates. The cyan and magenta lines are the ww3 model results of surface waves at the positions of two EM-APEX floats, respectively. The black dotted line in (d) indicates $\Delta\theta_p = 60^\circ$, the criterion suggested by Thomson et al. (2018) for identifying reliable wave direction estimates.

the storm-induced surface wave field extensively (Moon et al. 2004; Fan et al. 2009; Fan and Rogers 2016; Liu et al. 2017).

The estimated H_s at two floats is similar, about 2 m at 0800 UTC 16 October, and then increases to the peak of 10 m when Megi's eye passes the floats (Fig. 8). The simulated H_s in the ww3 is mostly within 1 m lower than the observations. Unlike the observed H_s , the simulated H_s at em4913a is higher than that at em3766c near the eyewall. Within the zonal distance of 100 km to the eye of Megi ($|X| < 100$ km), the f_p in the model results is slightly lower/higher than that at em3766c/em4913a.

The observed f_p may have more significant spatial variation than the H_s under Megi, different from the model simulations. Note that the ww3 simulation of surface waves in this study is initialized without the background surface wave field. It may bias the simulated wave frequency when the $|X|$ is still larger than 100 km.

The propagation direction and spread of dominant waves $\bar{\theta}_p$ and $\Delta\bar{\theta}_p$ at the f_p are then discussed (Fig. 8). Because the f_p varies significantly in the bimodal spectrum of $E(f)$, the results of $\bar{\theta}_p$ and $\Delta\bar{\theta}_p$ will not be discussed when the standard deviation

of f_p is >0.02 Hz. At em4913a, the $\bar{\theta}_p$ is between 120° and 150° at $X < -250$ km, and then rotates clockwise until Megi passes the floats. The $\Delta\theta_p$ decreases with the decreasing $|X|$ before the passage of Megi. Dominant waves near the eyewall may have less spread in direction due to the alignment between the wind and surface waves (Fig. 7). The difference of $\bar{\theta}_p$ between the observations and model results is less than 30° near the eyewall. After the passage of Megi's eye, the $\bar{\theta}_p$ at em4913a rotates to 120° , and the $\Delta\theta_p$ increases to $>60^\circ$. According to the spread of surface wave propagation direction $\Delta\bar{\theta}_p$ at the rear-right quadrant of Megi, waves at em4913a may be affected by the swell from the rear-left quadrant more than that at em3766c.

Except the EM-APEX float measurements during the ITOP experiment, a buoy is deployed for measuring surface waves at 127.25°E and 19.63°N since August 2010 (Collins et al. 2014; Collins 2014). The closest distance between the buoy and Megi's eye is ~ 100 km, at about half day after the passage to the floats. Because of different locations, the properties of surface waves measured by the floats cannot be directly compared to the buoy measurements. We therefore compare the ww3 model results to the observations at the buoy and floats (supplemental material, section B), respectively. When the $X < -100$ km, the ww3 fails to simulate the propagated swell at the buoy, the same as that found in the comparison between the float observations and model results. Within the $|X| < 100$ km, the simulated H_s , f_p , and θ_p are in good agreement with the wave measurements at two different platforms. The ww3 model results validated by the traditional wave measurements may indirectly support the robustness of float-measured surface waves. Future field experiments will be conducted for intercomparing the measurements between the floats and buoys directly.

c. Discussion on the wave spectrum estimation using limited measurements

Surface wave directional spectra are computed by using the measurements taken by the subsurface EM-APEX floats for the first time. The evolution of float-measured surface waves is in good agreement with the previous studies, no matter in the field experiments or model simulations (Wright et al. 2001; Moon et al. 2004; Chen et al. 2013). That is, EM-APEX floats can be used for exploring the interaction between the upper ocean structure and wind-induced surface waves under extreme environments. Note that setting $C_{\max} = 10$ will exclude spectral estimates at the high-frequency bands. The upper frequency limit of float-measured $\bar{E}(f)$ at all profiles is less than 0.2 Hz, even when the measurements in the upper 30 m are available. One may suspect the future application of EM-APEX floats for measuring high-frequency wind waves.

D'Asaro (2015) estimates the wave energy spectra $E(f)$ using the continuous float profiles in the upper 80 m (D'Asaro et al. 2014a). The individual estimates of $E(f)$ are the mean of the independent spectrograms over the period of one day, nearly the same as the traditional wave measurements at $f = 0.2$ – 0.5 Hz. Using measurements from multiple float profiles near the sea surface can reduce the uncertainty of estimated $E(f, \theta)$, assuming wave spectra changing insignificantly over this period. Therefore, future studies can still estimate the high-frequency wind waves by using the subsurface EM-APEX

floats, depending on the vertical trajectories and the length of the period for estimating wave spectra.

Unfortunately, because of the rapid growth of storm-induced wind waves, this study has no choice but to use the measurements in every single profile. Each window contains more than 90% overlapped measurements for computing the spectrograms, which may result in an artificial correlation between adjacent windows. The sensitivity test on the variability of the percentage of overlapped measurements is therefore performed (supplemental material, section D). The estimated $\bar{E}(f)$ around the f_p is similar with those using the windows with $\leq 50\%$ overlapped measurements. We may conclude that the properties of dominant waves under Megi are still captured by the floats reliably. It is no doubt that the potential uncertainty due to overlapped measurements can be further reduced if more measurements can be parsed into independent windows for the estimations. Other factors to the uncertainties of float-measured spectra will be explored by using the simulated measurements in the next section.

6. Uncertainties of wave estimates using simulated float measurements

Various factors can result in the uncertainties of float-measured spectra, such as the cutoff frequency f_c in the Butterworth high-pass filter and limited measurements in a single profile. In the following, the uncertainties will be explored by using simulated float measurements (supplemental material, section B), assuming the input of a surface wave energy spectrum E_{ref} in the form of the JONSWAP spectrum (Hasselmann et al. 1973).

a. Cutoff frequency f_c in the estimates of E_H

Different Ω is used for simulating the wave-induced electric current measured by the rotating electrodes, assuming the constant $W = 0.1 \text{ m s}^{-1}$ and $\theta_{\text{ref}} = 120^\circ$. The E_H is estimated by using the simulated voltage measurements (Fig. 9) in the window of $T = 300$ s (with 290 s overlapped), so the discrepancy between the E_H and E_{ref} due to insufficient measurements can be minimized (supplemental material, section C).

The simulated voltage measurements ϵ_i [Eq. (2)] are high-pass filtered by using the constant f_c ($=0.15$ Hz) and varied f_c ($=f_0 + \Omega/2\pi$ Hz), respectively. Note that the frequency of the low-frequency portion of ϵ_i (the frequency difference between waves and Ω) increases with the increasing Ω . For the ϵ_i measured by the electrodes with faster Ω , less variance of ϵ_i is filtered out by using a constant f_c , so the E'_H [Eq. (9)] will be overestimated more. Adjusting f_c based on the Ω is thus required. Compared with the E_H estimated using the low f_0 , the E_H computed using the high f_0 slightly overestimates/underestimates the spectral level at the low/high-frequency bands, but the effect of f_0 on $\bar{E}_H$ around the f_p is negligible. Therefore, $f_0 = 0.1$ Hz is used in this study in the varied f_c for computing the E_H .

b. Limited measurements for computing spectrograms

The apparent spectrograms E'_H and E'_Z are estimated by using the simulated float measurements, assuming the constant $W = 0.1 \text{ m s}^{-1}$, $\theta_{\text{ref}} = 120^\circ$, and $\Omega/2\pi = 0.1$ Hz. The simulated

FIG. 9. Results of adjusted spectrograms E_H assuming (a) constant cutoff frequency $f_c = 0.15$ Hz or (b)–(d) varied $f_c = \Omega/2\pi + f_0$ in the second-order Butterworth high-pass filter [with different f_0 in (b)–(d)], estimated by using the simulated float measurements of surface waves taken by the different rotation rate of EM sensors Ω (colored lines). The black lines are the input of a surface wave spectrum E_{ref} for constructing the simulated surface wave field.

measurements are parsed into the moving windows of $T = 120$ s, the same as those under Megi. The adjusted spectrograms E_H and E_Z are first derived by using the $c = 0$ in Eq. (A1) for correcting E'_H and E'_Z (Figs. 10a,b). The discrepancy of E_H and E_Z from the E_{ref} increases with the increasing Z . But, using measurements with a longer window of $T = 300$ s can reduce the biases significantly (supplemental material, section C). We suspect that the Z in the windows of short T may not be able to correct the apparent spectrograms reliably. The mean ratio of adjusted spectrograms to E_{ref} within the frequency band from 0.07 to 0.14 Hz is computed using different values of c (supplemental material, section C). The estimates of $\bar{E}_H$ and $\bar{E}_Z$ using the $c = 0.06$ and 0.09, respectively, are nearly the same as the E_{ref} (Figs. 10c,d).

The uncertainty due to W is further studied (Figs. 10c,d). Float measurements are simulated assuming a constant $\Omega = 0.1$ Hz. The effect of W on the spectral estimates around the f_p is insignificant, justified by the negligible discrepancy between the adjusted spectrograms and E_{ref} . Because high-frequency surface waves decay more rapidly in depth than low-frequency waves, fewer measurements can be taken by the fast-moving floats near the sea surface and used for estimating the spectral level at the high-frequency bands. The upper frequency bound of $\bar{E}_H$ and $\bar{E}_Z$ therefore decreases with the increasing W . Fortunately, at least four spectrograms have estimates up to 0.2 Hz at $W < 0.18$ m s⁻¹. Considering the f_p of surface waves under tropical cyclones is mostly 0.08–0.1 Hz, a single profile of wave measurements on EM-APEX floats ($W \sim 0.11$ m s⁻¹) would be sufficient for estimating surface wave spectra reliably.

c. Unknown noise during profiling

Compared with the measured low-frequency waves, measurements of high-frequency surface waves may be biased by

the noise at the deeper layers more due to different depth-decaying rate. Because it is difficult to quantify the biases associated with the unknown noise accurately, this study is aimed to discuss the uncertainties due to the white noise. The simulated float measurements (as that in section 6b) plus the white noise (δ_{eh} is added to the measurements on E_1 in the form of velocity; δ_{az} is added to the vertical acceleration measurements) will be used for estimating the E_H and E_Z .

Spectral estimates using the measurements without the white noise are reliable at $f < 0.12$ Hz, even the median depth Z of spectrograms is 70 m (Fig. 11). Once the white noise contaminates the float measurements, its effect on the spectral estimates is significant, especially for the spectrograms at the deep layers. Therefore, increasing the amplitude of δ_{eh} or δ_{az} will bias more spectral estimates at the high-frequency bands. These unrealistically amplified spectral estimates are caused by the correction on the depth-decaying effect, when the variance of surface waves at the high-frequency bands is smaller than the white noise. Because the C_{max} is used for excluding the spectral estimates at $f > \sqrt{g \log(C_{\text{max}})/[(2\pi)^2 Z]}$ (section 3), choosing the value of C_{max} is important for removing those spectral estimates that have been unrealistically amplified by the white noise.

The effect of C_{max} is studied (Fig. 11). Without the white noise, $C_{\text{max}} = 90$ can be used for correcting the depth-decaying effect at all spectrograms whose $Z < 70$ m. If the δ_{eh} increases to 0.02 m s⁻¹, which is similar with the upper limit of instrumental noise of float velocity measurements (Hsu et al. 2017), $C_{\text{max}} = 30$ is used for estimating E_H . The noise in the vertical acceleration measurements is ~ 0.003 m s⁻² (<https://www.analog.com/media/en/technical-documentation/obsolete-data-sheets/ADXL311.pdf>), much smaller than the noise used in the

FIG. 10. Estimates of adjusted spectrograms (a) E_H and (b) E_Z at different median depths of the spectrograms Z assuming $c = 0$ in Eq. (A1), and estimated ensemble mean of (c) $\overline{E_H}$ and (d) $\overline{E_Z}$ by using different vertical profiling speed of the floats W , assuming $c = 0.06$ and 0.09 , respectively. The black lines are the input of a surface wave spectrum E_{ref} for constructing the simulated surface wave field.

simulated δ_{az} . That is, $C_{max} = 10$ can be used for computing the adjusted spectrograms at EM-APEX floats reliably, the same as that used in D'Asaro (2015).

7. Summary and conclusions

Two pairs of orthogonal electrodes E_1 and E_2 mounted on the subsurface EM-APEX floats are used for measuring the electric current induced by the horizontal motion of seawater as the voltage. Unlike the low-frequency (<0.02 Hz) voltage measurements which have been extensively used for computing the velocity of ocean current (e.g., Sanford et al. 2005; Hsu et al. 2017), the high-frequency voltage measurements induced by the horizontal motion of surface waves are often treated as the residuals. Therefore, this study proposes a new method for deriving the horizontal velocity of surface waves. Combined with the measured vertical acceleration of surface waves, the surface wave directional spectrum $E(f, \theta)$ can be directly estimated using float measurements.

The $E(f, \theta)$ is estimated using EM-APEX float measurements under a Tropical Cyclone Megi (D'Asaro et al. 2014b) for the first time. We exclude the spectrograms measured by the slow-moving floats, and the spectrograms that may have

been biased by the data processing. Though the upper bound of frequency of $\overline{E}(f)$ is limited to below 0.13 Hz due to the missing measurements in the upper 30 m after 1400 UTC 16 October 2010, the peak frequency $f_p \sim 0.08$ –0.1 Hz near the eyewall of Megi is captured by float measurements. The check factor R as the ratio of E_H to E_Z , is about one at the peak frequency f_p in most profiles, consistent with the linear wave theory. The uncertainties of surface wave estimates are also studied using simulated float measurements. E_H and E_Z estimated using the EM-APEX float measurements are reliable, despite that no other wave measurements are available for the intercomparison.

When Megi moves closer to the floats, the wind rotates clockwise and becomes more aligned with the direction of dominant waves. The extended fetch effect (Young 2006) on the right of Megi's track enhances the H_s to >10 m and downshifts the f_p to 0.08 Hz. After the passage of Megi's eye to the floats, the f_p increases to 0.1 Hz. Though the $\overline{E}(f)$ is still a monomodal spectrum in frequency, surface waves at 43 km to the right of Megi's track appear to propagate at three different directions at the frequency bands from 0.09 to 0.11 Hz. Surface waves at the rear-right quadrant of Megi may be affected by the swell from the rear-left quadrant. Overall, the spatial distribution of the angle between the dominant wave direction

FIG. 11. Estimates of adjusted spectrograms (black solid lines) (a)–(c) E_H and (d)–(f) E_Z at different median depths of the spectrograms Z (y axis) by using the simulated float measurements with different white noise. The noises of horizontal velocity measurements δ_{eh} and vertical acceleration measurements δ_{az} are added to the measurements in (a)–(c) and (d)–(f), respectively. The colored lines with different values of C_{max} indicate their upper limits of reliable spectral estimates. The black dashed lines are the input of a surface wave spectrum E_{ref} for constructing the simulated surface wave field.

and wind is in good agreement with the results from different previous studies (e.g., Wright et al. 2001; Moon et al. 2004; Chen et al. 2013). The measured dominant wave direction can benefit the parameterization of wind forcing on the surface wave field in the models (Donelan et al. 2012) or momentum transfer efficiency (Hsu et al. 2019).

This study proposes a new method for estimating surface wave directional spectra $E(f, \theta)$ using the subsurface EM-APEX float measurements. Because temperature, salinity,

current velocity, and $E(f, \theta)$ can be measured simultaneously, the roles of surface waves in the evolution of upper ocean structure can be better studied using float measurements in the future. For studies interested in estimating waves at >0.2 Hz, profiling floats near the sea surface will be useful, such as from surface to 20-m depth. Measuring the evolution and turbulent mixing in the surface mixed layer (i.e., in the upper 100 m) is also critical for quantifying the effects of surface waves. Here, we have described the details for computing $E(f, \theta)$ using

measurements from EM-APEX floats. Future work will focus on the intercomparison between the float-measured surface waves and wave measurements taken by other platforms.

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APPENDIX

Method for Correcting the Apparent Spectrograms on Subsurface Floats

Estimates of surface wave spectra using measurements taken by subsurface floats can include biases, mostly due to the decaying amplitude of surface waves in depth. D’Asaro (2015) therefore introduces a series of steps for correcting the biases, including the depth correction, vertical motion correction, sampling interval correction, and spectral spreading correction. Because estimating the noise in the individual spectrograms is difficult, this study chooses to quantify the effect of unknown noise on the estimated spectrograms (section 6), instead of subtracting the spectral level of white noise from the estimated spectra, different from D’Asaro (2015). The spectrograms derived directly from the autospectra of horizontal or vertical motion of surface waves are termed apparent spectrograms $E^{(0)}$ (the number in the superscript represents the steps). The adjusted spectrograms will be derived after correcting all biases in $E^{(0)}$.

Because surface waves decay exponentially with depth (termed the depth-decaying effect), the estimated spectral level in $E^{(0)}$ is corrected by multiplying the exponential decaying rate at each frequency band, i.e.,

$$E^{(1)} = E^{(0)} e^{2kZ} \quad \text{for } e^{kZ} < C_{\max},$$

where Z is the median depth of each spectrogram, positive downward. A parameter $C_{\max}$ is defined for excluding estimates at $e^{kZ} > C_{\max}$, because the e^{2kZ} may unrealistically amplify the signals of high-frequency surface waves.

However, the corrected spectrograms at the deeper layers may still be underestimated if measurements in the windows are insufficient (supplemental material, section C). We suspect that the estimated median depth of the measurements Z in the windows of short T is unreliable. Compared to the method in D’Asaro (2015), an additional parameter c is added for the correction as

$$E^{(1)} = E^{(0)} e^{2k[1+c e^{(-T/T_0+1)}]Z} \quad \text{for } e^{kZ} < C_{\max}, \quad (\text{A1})$$

where $T_0 = 120$ s. We explore the values of c in supplemental material, section C, suggesting that $c = 0.06$ and 0.09 are the most reliable values for correcting the apparent spectrograms E'_H and E'_Z , respectively.

Because the measurements within each window also vary in depth, fast W will result in more variation of wave amplitude. It will bias the spectral estimates at the high-frequency bands. The vertical motion correction (D’Asaro 2015) on the spectrograms estimated using the taper of Hanning window can be performed as

$$E^{(2)} = E^{(1)} \left[\frac{(\lambda/2)^2 + \pi^2}{\pi^2} \frac{(\lambda/2)}{\sinh(\lambda/2)} \right]^2, \quad (\text{A2})$$

where $\lambda = kWT$, and W is the vertical speed of the float.

Because the data taken by the floats are output in 1 Hz, the discrete measurements in the interval of 1 s will slightly affect the spectral level at the high-frequency bands. The $E^{(2)}$ can be corrected by using the quadratic polynomial in D’Asaro (2015), i.e.,

$$E^{(3)} = E^{(2)} / (d_1 \lambda^2 + d_2 \lambda + d_3), \quad (\text{A3})$$

where $d_1 = 0.004897$, $d_2 = 0.033609$, and $d_3 = 0.999897$ for the Hanning window, respectively. Finally, surface waves at two sidebands of the Hanning window have different depth-decaying rate, termed spectral spreading correction. The $E^{(4)}$ (also termed the adjusted spectrograms) can be derived after the correction as

$$E^{(4)} = E^{(3)} \left\{ 1 + 2h \left[\cosh \left(\frac{2\omega\delta\omega + \delta\omega^2}{g} Z \right) - 1 \right] \right\}^{-2}, \quad (\text{A4})$$

where $\omega = 2\pi f$ is the angular frequency and $\delta\omega = 2\pi/T$.

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