On the Wave State Dependence of the Sea Surface Roughness at Moderate Wind Speeds under Mixed Wave Conditions

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(Manuscript received 13 May 2020, in final form 19 July 2020)

ABSTRACT: Wind stress depends on the sea surface roughness, which can be significantly changed by surface wind waves. Based on observations from a fixed platform, we examined the dependences of the sea surface roughness length on dominant wave characteristic parameters (wave age, wave steepness) at moderate wind speeds and under mixed-wave conditions. No obvious trend was found in the wave steepness dependence of sea surface roughness, but a wave steepness threshold behavior was readily identified in the wave age dependence of sea surface roughness. The influence of dominant wind waves on the surface roughness was illustrated using a wind–wave coupling model. The wave steepness threshold behavior is assumed to be related to the onset of dominant wave breaking. The important role of the interaction between swell and wind wave was highlighted, as swell can absorb energy from locally generated wind wave, which subsequently reduces the wave steepness and the probability of dominant wave breaking.

KEYWORDS: Wind stress; Wind waves; Air–sea interaction; Parameterization

1. Introduction

Wind stress over the sea surface is the primary driving force for upper-ocean dynamics, including circulation and waves. An accurate estimate of wind stress is important for modeling oceanic and atmospheric interactions and their respective dynamics. In modeling studies, the wind stress is generally calculated from the so-called bulk formulation as follows:

$$\tau = \rho_a C_D U_{10}^2, \quad (1)$$

where U_{10} is the wind speed at a height of 10 m, C_D is the drag coefficient, and ρ_a is the air density. Following Monin–Obukhov similarity theory, the neutral drag coefficient is equivalent to the sea surface roughness length z_0 :

$$C_{DN} = \kappa^2 \ln^{-2} \left(\frac{10}{z_0} \right), \quad (2)$$

where κ is the von Kármán constant, and the subscript N denotes neutral conditions. Substantial efforts have been devoted to the parameterization of the drag coefficient and the sea surface roughness length, in which the wind speed is frequently used as a candidate. The classic parameterization of the sea surface roughness length was proposed by Charnock (1955):

$$\frac{gz_0}{u_*^2} = \alpha, \quad (3)$$

where u_* is the friction velocity related to the wind stress $\tau = \rho_a u_*^2$, g is the gravitational constant, and α is known as the Charnock parameter, which was found to vary significantly from 0.012 to 0.035 (e.g., Garratt 1977; Wu 1980; Guan and Xie 2004).

The impact of the wind–wave interaction on wind stress has been widely documented in both theoretical and observational studies (e.g., Donelan et al. 1993; Rieder et al. 1996; Hara and Belcher 2002). The wave age, which describes the development state of local wind waves, has been preferentially used in the parameterization of α in several studies (e.g., Johnson et al. 1998; Jones and Toba 2001; Oost et al. 2002; Gao et al. 2006). However, different experiments have shown considerable changes in the wave-age dependence of α , and even contradictory results have been reported between field and laboratory measurements (Toba et al. 1990; Donelan 1990).

Scaling α with the wave age may suffer from spurious self-correlation, as both parameters depend on the friction velocity. Alternatively, Hsu (1974) proposed a dimensionless value of z_0 normalized by the significant wave height H_s . Subsequently, efforts have been devoted to relating z_0 to the wave age (Smith et al. 1992; Donelan et al. 1993; Drennan et al. 2003, hereafter D03),

$$\frac{z_0}{H_s} = A_1 \left(\frac{u_*}{c_p} \right)^{B_1}, \quad (4)$$

where c_p is the phase velocity of the dominant wind wave. Note that the variability in u_*/c_p is dominated by variability in u_* , D03 sought to solving this problem by combining data from many field experiments and grouping the data by u_* , their results gave $A_1 = 3.35$ and $B_1 = 3.4$.

Denotes content that is immediately available upon publication as open access.

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DOI: 10.1175/JPO-D-20-0102.1

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In place of wave age, some efforts have been devoted to using other wind-wave characteristic parameters in the parameterization of roughness. Taylor and Yelland (2001, hereafter TY01) used the wave steepness as a candidate, and they proposed the following relation for sea states ranging from strongly forced to shoaling,

$$\frac{z_0}{H_s} = A_2 \left(\frac{H_s}{L_p} \right)^{B_2}, \quad (5)$$

where L_p is the dominant wavelength. The coefficients $A_2 = 1200$ and $B_2 = 4.5$. More recently, Takagaki et al. (2012, hereafter T12) derived $A_2 = 10.87$ and $B_2 = 3.0$ for sea states from normal to strong wind conditions.

Drennan et al. (2005) performed a comprehensive assessment of the performance of the wave-age- (D03) and wave-steepness-dependent (TY01) scaling using a combination of eight datasets representing a wide range of sea states. They concluded that both scalings yield improved estimates compared to a sea-state-independent scaling under wind-wave dominant conditions, but their performance degrades significantly with an increasing contribution of the swell component.

Swell either is generated elsewhere or results from the slowing of local winds or a change in direction. In the open ocean, waves are commonly mixed with wind waves and swells. Swells are typically energetic and even dominant, significantly changing the wave age and wave steepness compared to pure wind-wave conditions (Hanley et al. 2010; Semedo et al. 2011; Li et al. 2018a), but swells do not contribute as much to the sea surface roughness. García-Nava et al. (2012b) argued in support of using only the wind sea part in the parameterization of sea surface roughness. Hwang et al. (2011a) proposed using a characteristic wave frequency defined as a weighted average of swells and wind waves to parameterize the drag coefficient in mixed waves. Smedman et al. (2003) derived from observation analysis that the drag coefficient is governed by the wave energy spectral ratio between swell and wind-wave, as well as the wave age. Carlsson et al. (2010) suggested that the impact of swell on the wind-wave interaction can explain the discrepancies among the different investigations. The impact of swell on the wind-wave interaction was discussed by Hwang et al. (2011b), who suggested that the development of wind waves tends to be enhanced in the presence of swell, and by García-Nava et al. (2009, 2012a), who suggested that swell can reduce the wave-supported stress by suppressing the intensity of short wind waves in the equilibrium range. Nevertheless, the wind-wave interaction in the presence of swell is much less understood than the interaction for pure wind waves, largely accounting for the slow development of sea state scaling under mixed wave conditions.

The goal of this study is to address the impact of swell on the wind-wave interaction and the wave state dependence of sea surface roughness. First, we analyze the wave age and wave steepness dependencies of the sea surface roughness at moderate wind speed under open ocean conditions, where mixed waves frequently occur (section 3), using data observed from a fixed platform (section 2). Then, a wind-wave coupling model is used to analyze the physics behind the wave state

dependence, and the impact of swell is specifically addressed (section 4). Finally, the discussions and conclusions are presented in section 5.

2. Observations

a. Data

The measurements were taken from the Flux Observation Project in the South China Sea (FOPSCS), which was conducted in the coastal waters of the northern South China Sea (Fig. 1). An eddy covariance system was configured to measure the momentum, heat and gas flux across the air-sea interface, including an R.M. Young sonic anemometer mounted at a height of ~ 19 m above the sea surface, which was used to measure the wind velocity and sonic temperature at a sampling frequency of 10 Hz, and an open-path $\text{CO}_2/\text{H}_2\text{O}$ analyzer operated synchronously to observe the water vapor concentration at the same sampling frequency. The wind stress can be directly determined from the measured wind fluctuations using the eddy covariance method, where the average time scale is determined hourly by the multiresolution decomposition method (Vickers and Mahrt 2003). A series of data processing and quality control procedures, including spike removal and tilt correction, were applied to determine the flux, as detailed in our previous studies (Li and Zhao 2016; Zou et al. 2017). Flux observations with winds blowing from the northwest were discarded as they may have been distorted by the platform. Four wind anemometers were mounted at heights of 31.3, 23.4, 20.0, and 16.4 m above mean sea level to measure the hourly averaged wind speed and direction.

A Nortek acoustic wave and current profiler (AWAC) was deployed on the seabed near the platform. The water depth was 15 m, which is considerably less than the maximum working depth (~ 60 m) of the AWAC. The frequency-directional spectrum $S(\omega, \theta)$ was estimated from the composition of a frequency spectrum $S(\omega)$ and a directional distribution at each frequency $D(\omega, \theta)$:

$$S(\omega, \theta) = S(\omega)D(\omega, \theta), \quad (6)$$

where ω is the radian frequency, and θ is the wave direction. The surface elevation detected via acoustic surface tracking (AST), was used to determine the frequency spectrum $S(\omega)$ using the Fourier transform method. A vertical beam was dedicated to the AST. The surface elevation was measured at an interval of 3 h with a sampling frequency of 2 Hz and a duration of 1024 s. Note that short waves with a frequency > 0.4 Hz may be contaminated by Doppler modulation (Banner 1990); following Zou et al. (2018), a spectral tail was added, with f^{-4} in the equilibrium range (up to $3f_p$, where f_p is the peak frequency) and f^{-5} in the saturation range. The directional distribution $D(\omega, \theta)$ was estimated from the surface elevation and velocity measurements using the SUV (referring to surface tracking, and orbital U and V velocities) method (Pedersen et al. 2005). To calculate the directional distribution of waves with the SUV method, information is necessary on the wave orbital velocities near the surface, and AST should be performed for accurate wave energy. Thus, the SUV method is

FIG. 1. Location of the platform.

similar to the classic PUV (pressure and U and V velocities) method, which is a triplet-type direction solution. Finally, the directional distribution can be derived by the maximum likelihood method. The wave directional estimates were indicated to be in good agreement with the surface wave buoy measurements (Pedersen and Siegel 2008).

b. Processing method

A two-dimensional spectral partitioning and identification method was used to separate the wind-wave and swell components. Assuming that waves comprise a mix with several independent wave systems, the spectral partitioning method was used to separate the wave systems using the method proposed by Hanson and Phillips (2001). A partitioned wave system was identified as a wind-wave component based on a wave age criterion:

$$\eta \frac{U_{10}}{c_p} \cos(\theta_p - \psi) \geq 1 \quad \text{and} \quad |\theta_p - \psi| \leq \frac{\pi}{2}, \quad (7)$$

where c_p is the phase speed of the spectral peak, θ_p is the corresponding wave direction, ψ is the wind direction, and η is a constant taken to be 1.3 (Hasselmann et al. 1996; Voorrips et al. 1997). Otherwise, a partitioned wave system was identified as a swell component as follows:

$$\eta \frac{U_{10}}{c_p} \cos(\theta_p - \psi) < 1 \quad |\theta_p - \psi| > \frac{\pi}{2}. \quad (8)$$

The surface roughness was not directly measured but could be estimated from the wind and flux measurements (e.g., D03; Drennan et al. 2005) by assuming a logarithmic profile shape of the wind speed U (Monin and Yaglom 1971):

$$U(z) = (u_* / \kappa) [\ln(z/z_0) - \phi_m(z/L)], \quad (9)$$

where z is the height above the sea surface, $L = -u_*^3 \theta_v / g \kappa w' \theta'_v$ is the Obukhov length scale, θ_v is the virtual temperature, $w' \theta'_v$

is the flux of the virtual temperature (or buoyancy flux), and ϕ_m is the atmospheric stability function. Atmospheric stability has a strong impact on the observed wind speed at a given height (Young and Donelan 2018). The boundary layer is unstable ($L < 0$) when the water is warmer than the air, and stable ($L > 0$) for colder water. The wind speeds have a more uniform vertical distribution with an unstable boundary layer. To exclude the effect of atmospheric stability, the wind speed is converted into the neutral wind speed (U_{10N}) at 10 m:

$$U_{10N} = U(z)(u_* / \kappa) [\ln(z/10) - \phi_m]. \quad (10)$$

Hereafter, U_{10N} is simplified to be U_{10} for convenience. The atmospheric stability function ϕ_m is used in the model developed by Höglström (1988) because it fits well with the observed results from the FOPSCS (Zou et al. 2017).

The sea surface roughness can be directly calculated from the neutral wind speed and friction velocity by the following:

$$z_0 = 10 \exp[-U_{10} \kappa / u_*]. \quad (11)$$

A logarithmic wind profile over wind waves has been supported by observations taken from a land-based tower (e.g., Smedman et al. 2003; Höglström et al. 2008). However, the wind profile can be disturbed by a swell-induced momentum flux in a boundary layer over a wavy surface (Babanin et al. 2018; Zou et al. 2018). Many studies (e.g., Grare et al. 2013; Kahma et al. 2016) have shown that a swell can exert a distinct peak in wind spectra centered at the frequency of swell waves, as a result, the behavior of the air-sea interaction is different from that of the interaction under wind waves. We recently found (Zou et al. 2018) that the swell-induced flux estimated using the wave measurements from the FOPSCS is significant at low wind speeds but accounts only for less than 5% of the total wind stress for $U_{10} > \sim 4 \text{ m s}^{-1}$. We compared the logarithmic profile with the wind profile measurements for $U_{10} > 4 \text{ m s}^{-1}$ (Fig. 2), and the results show that the observed mean wind profile is well reproduced by the logarithmic profile.

FIG. 2. Mean wind profile (stars) derived from the time-averaged normalized wind speed at different heights and the curve (blue line) fitted to the logarithmic profile of Eq. (9) using a regression method.

Therefore, we restricted the analysis to this wind range. Finally, a number of 182 coincident wind and wave data pairs was derived.

c. Wind and wave conditions

The temporal variation in the wind and wave conditions are illustrated in Fig. 3, and the statistics were presented in Fig. 4.

Figure 3a shows the variation in the wind speed (black stars and left-hand scale) and the significant wave height (blue circles and right-hand scale) during the observation time period. The winds generally lie in the range of $4\text{--}10\text{ m s}^{-1}$, representing moderate wind speeds that center at approximately 7 m s^{-1} (Fig. 4a). The significant wave height is in the range of $0.5\text{--}1.7\text{ m}$. The swell index is estimated as the energy ratio between swell and wind wave, and ranges from 0 to 5 (Fig. 4b). The wave age, defined as the ratio of the phase velocity of the dominant wind wave to the friction velocity, covering a wide range from 9 to 35 (Fig. 4c), but the wave ages are dominated by relatively large values (>15) representing well-developed wind waves.

Figure 3b shows the directions of the winds (black stars) and dominant waves (spectral peaks, blue circles); the directions signify where the winds and waves from and the convention is clockwise relative to north. The winds generally come from the east with a mean azimuth of 107.5° , and the azimuths of the dominant waves are slightly larger with a mean value of 126.7° . It was found that $\sim 63\%$ of the spectral peaks are occupied by the swell components. The deflection angles between the wind waves and swells mostly lie in the range of $0^\circ\text{--}30^\circ$ (Fig. 4d), representing the predominance of following-swell conditions.

Figure 3c shows the phase speeds c_0 of dominant waves and wind waves. The phase speed is calculated as $c_0 = \omega_p/k_p$, in which the subscript p denotes the spectral peak, and k is the wavenumber, which can be calculated interactively from the dispersion relation $\omega_p^2 = gk_p \tanh(k_p h)$, where h is the water depth. To check for the possible shoaling effect in nearshore regions (Smedman et al. 1999), the phase speeds of the

FIG. 3. (a) Hourly mean wind speed U_{10} (black crosses) and significant wave height (blue circles) from January 2012 to July 2012; (b) wind direction (black crosses) and dominant wave direction (blue circles), where the directions signify where the winds and waves are from and the convention is clockwise relative to north; (c) phase speeds of dominant waves (black crosses) and wind waves (green crosses). Deep-water values are denoted by circles (blue for dominant waves and red for wind waves).

FIG. 4. Relevant wind and wave statistics in the dataset: Probability distribution of the (a) wind speed, (b) swell index, (c) wave age, and (d) deflection direction between wind waves and swells.

dominant waves and wind waves in deep water ($c_0 = g/\omega_p$) are shown for comparison. As seen, when the phase speeds of the dominant waves approximate 11 m s^{-1} , they are largely determined by the depth and clearly deviate from the deep-water values, with $\sim 60\%$ of the cases having a difference from 10% of the deep-water value, while the phase speeds of wind waves mostly equal the deep-water values (only $\sim 6\%$ of cases have a difference larger than 10% from the deep-water values). These results indicate that swells frequently experienced the shoaling effect, while wind waves are minimally influenced by the shallow water depth. Since we focus on the wind-wave dependence of sea surface roughness, the shoaling effect should not have a direct impact on the results.

3. The wave state dependence of the sea surface roughness

The wave state dependence of the normalized roughness length z_0/H_s is given in Fig. 5. The wave age and wave steepness scalings were overlain for comparison. The bin-averaged data from Drennan et al. (2005) are also shown for reference. Note that the wave state dependence of sea surface roughness was investigated using either the data of the wave parameters corresponding to the wind-wave spectrum (e.g., D03) or those corresponding to the full spectrum (e.g., TY01). For the purpose of this test, these two choices were used separately. To check the wave-state-dependent scalings

under different wave conditions, we partitioned the wave conditions according to the swell index, with values < 1 indicating wind-wave dominance and the remaining values indicating mixed wave condition.

For the wave steepness dependence, no clear trend can be identified when either the full spectrum or the wind-wave spectrum is considered (Figs. 5a,b), and the correlation coefficients are -0.03 and 0.09 , respectively. The correlations are much higher when the wave age is considered (Figs. 5c,d): the correlation coefficient is 0.47 when the full spectrum is considered, and reaches 0.89 when the wind-wave spectrum is considered. Notable scatters are observed at a given wave age or wave steepness. Uncertainties may exist in discriminating wind waves and swells, Li and Zhao (2012) suggested that separated wind waves can be underestimated by up to $\sim 6\%$, which would result in an overestimation of z_0/H_s by up to $\sim 12\%$ in the wave steepness dependence and $\sim 6\%$ in the wave age dependence.

Drennan et al. (2005) proposed a wave steepness threshold of $H_s/L_p = 0.02$, and they suggested that the sea-state scalings (D03, TY01) hold for $H_s/L_p \geq 0.02$. Since the wave steepness dependence is quite weak, this dependence cannot be readily described by either TY01 or T12. This may be because the observed H_s/L_p values are generally small with values generally less than 0.04 , while TY01 and T12 are both developed for much steeper waves. The wave age dependence can be

FIG. 5. Wave state dependence of the normalized z_0 : (a) wave steepness dependence using the full spectrum; (b) wave steepness dependence using the wind-wave spectrum, where the vertical dashed lines denote $H_s/L_p = 0.02$; (c) wave age dependence using the full spectrum; (d) wave age dependence using the wind-sea spectrum property. The bin-averaged data from Drennan et al. (2005) are overlain for reference (denoted by circles). The lines represent the proposed sea-state-dependent scalings. The error bars represent 60% of the standard deviation in a bin.

effectively described by D03 for $H_s/L_p \geq 0.02$ when only the wind-wave spectrum is considered but is systematically underestimated by D03 for $H_s/L_p < 0.02$.

4. Physical analysis

a. Wind-wave interaction

To address the physics behind wave state dependencies, we performed an analysis with a wind-wave coupling model (Kudryavtsev and Makin 2007; Kudryavtsev et al. 2014). The wind stress at the sea surface contains the viscous stress and the form drag in the presence of surface gravity wave:

$$\tau = \tau_f + \tau_v. \quad (12)$$

The viscous stress has also been called the “skin” drag, which arises from direct molecular interactions in the molecular sublayer (Kudryavtsev and Makin 2001):

$$\tau_v = \frac{1}{\kappa d} \ln \left(\frac{dv}{z_0 u_*} \right) u_*^2, \quad (13)$$

where v is the molecular viscosity and $d = 10$ is the molecular sublayer constant. The form stress τ_f is contributed to by both breaking and nonbreaking waves, but in quite different ways: nonbreaking waves support the growth of wind waves by absorbing energy from wind, resulting in wave-induced stress, while breaking wind waves cause airflow separation (AFS) at the crest, generating AFS stress (Kudryavtsev and Makin 2001; Kukulka and Hara 2008a,b). As a result, the form drag is expressed as a sum of the wave-induced stress τ_w and AFS stress τ_s in the spectral domain:

$$d\tau_f(\mathbf{k}) = d\tau_w(\mathbf{k}) + d\tau_s(\mathbf{k}), \quad (14)$$

where

FIG. 6. Comparisons between the friction velocity observed on the platform and those estimated from the analytical model. The error bars represent the standard deviations of the mean.

$$\begin{aligned} d\tau_w(\mathbf{k}) &= c_\beta u_*^2 \cos^3 \theta k^{-2} B(\mathbf{k}) d\mathbf{k}, \\ d\tau_s(\mathbf{k}) &= c_s u_*^2 \cos^3 \theta k^{-1} \Lambda(\mathbf{k}) d\mathbf{k}, \end{aligned} \quad (15)$$

where k is the wavenumber, θ is the wave direction, c_β is the growth rate coefficient, c_s is the separation stress coefficient, $B(\mathbf{k})$ is the saturation spectrum, $\Lambda(\mathbf{k})$ is the spectral density of the length of the wave breaking front (Phillips 1985), and the quantity $\Lambda(\mathbf{k})$ is related to $B(\mathbf{k})$ by assuming that the spectral energy loss due to wave breaking $D(\mathbf{k})$ is balanced with the wind energy input $I(\mathbf{k})$:

$$D(\mathbf{k}) = b g^{-1} c^5 \Lambda(\mathbf{k}) \simeq I(\mathbf{k}) = \beta \omega g k^{-4} B(\mathbf{k}), \quad (16)$$

where c is the phase speed, β is the wave age, and b is the effective breaking strength parameter, which spans in the range of 10^{-4} – 10^{-1} as reported by different experiments (Melville and Matusov 2002; Babanin and Young 2005; Thomson et al. 2009; Gemmrich et al. 2013), it was found the model results fit quite well with observations (as shown in Fig. 6) when b is taken to be 0.01.

Long breaking wind-wave fronts generate as AFS region, in which the form stress supported by short wind waves and the viscous stress locally vanish (Reul et al. 2008; Buckley and Veron 2016). Considering the sheltering effect, the wave-induced stress is reduced to $d\tau_{w\text{-sheltering}}(\mathbf{k}) = [(1 - q(k)) d\tau_w(\mathbf{k})]$, and the viscous stress is $\tau_{v\text{-sheltering}} = [(1 - q(k_b)) \tau_v]$, where $k_b \approx 2\pi/0.15 \text{ rad m}^{-1}$ is the wavenumber of the shortest breaking wind wave providing the AFS and q is the cumulative fraction of the sheltered surface describing the cumulative contribution of breaking sea waves to the sheltered zones and is determined from the local integration of $k^{-1} \Lambda(\mathbf{k}) d\mathbf{k}$. Because q has an upper limit of 1, the threshold wavenumber k_0 is introduced as $q(k_0) = 1$. Assuming that breaking wind waves with the wavenumber $k > k_0$ do not contribute to the total AFS stress as they are trapped in longer waves, the AFS stress is

expressed as $d\tau_{s\text{-sheltering}}(\mathbf{k}) = h(k - k_0) d\tau_s(\mathbf{k})$, where $h(x)$ is the Heaviside step function.

Finally, the total stress can be rewritten as follows:

$$\tau = \int d\tau_{w\text{-sheltering}}(\mathbf{k}) + \int_{k < k_b} d\tau_{s\text{-sheltering}}(\mathbf{k}) + \tau_{v\text{-sheltering}}. \quad (17)$$

The surface stress can be determined from the wind-wave coupling model via an iterative solution of Eqs. (12)–(17). The modeled surface stresses are compared with the measured momentum flux in Fig. 6 revealing a good agreement between them.

We plotted the normalized viscous stress and form drag separately against the wave steepness (Fig. 7). The viscous stress is generally larger than the form drag, particularly at low wave steepness. The viscous stress is clearly reduced by the AFS effect for $H_s/L_p \geq 0.02$, as inferred from the difference between the viscous stress with and without considering the AFS effect. The AFS effect strongly depends on the wave steepness: it is stronger in cases with higher wave steepness. The AFS effect is rather weak for $H_s/L_p < 0.02$. This phenomenon is expected because the wave steepness is frequently used to characterize the probability of wave breaking: wave breaking is more likely to occur for a higher wave steepness. Banner et al. (2000) found that the occurrence of dominant wave breaking is closely related to the so-called significant spectral peak steepness δ_p , defined as $\delta_p = (H_p k_p)/2$, where $H_p = 4 \left[\int_{0.7\omega_{pws}}^{1.3\omega_{pws}} S(\omega) d\omega \right]^{1/2}$ and ω_{pws} corresponds to the peak frequency of the wind wave in radian. It was suggested that dominant wave breaking does not occur up to a threshold value δ_p of approximately 0.05–0.06. The platform observations indicated that δ_p is closely correlated with H_s/L_p with a correlation coefficient of 0.93, and values in the range of $\delta_p = 0.05$ – 0.06 roughly correspond to H_s/L_p in the range of 0.02–0.025. It is very likely that the occurrence of dominant wind-wave breaking accounts for the wave steepness threshold behavior.

The form drag exhibits a generally increasing trend with the wave steepness. Wave-induced stress dominates the form drag and clearly increases with increasing wave steepness regardless of the increasing AFS effect. AFS stress generally increases with increasing wave steepness, and the trend appears to be more significant for large wave steepness ($H_s/L_p > 0.02$).

To summarize, for $H_s/L_p < 0.02$, the wave steepness dependence of the surface stress is mostly attributable to wave-induced stress, and the AFS effect is negligible, but the wave steepness dependence of the total drag is not significant because the form drag is much smaller than the viscous stress. For $H_s/L_p > 0.02$, a higher wave steepness corresponds to a higher form drag and a stronger AFS effect, the latter results in a lower viscous stress. As a result, the counteracting effect results in a relatively weaker dependence on the total surface stress than that on the form drag. This offers a possible explanation for the discrepancy between the observations and the two wave-steepness scalings (TY01; T12): these scalings were both developed for much steeper waves that typically occur under high winds or experience a strong shoaling effect, in which the form drag is dominant and the viscous stress is

FIG. 7. Normalized viscous stress (with and without considering the AFS effect for red and green crosses), the form drag (blue crosses) and the total surface stress (black crosses) of the model estimates vs H_s/L_p . VSA and VS denote the viscous stress with and without considering the AFS effect, respectively. FSA denotes the form drag considering the AFS effect, and TSS denotes the total surface stress. The data are averaged in bins of H_s/L_p .

rather weak (Kudryavtsev and Makin 2007; Mueller and Veron 2009), which may contribute to a stronger wave steepness dependence.

b. The swell effect

To further interpret the wind-wave interaction in mixed waves, it was necessary to address the wind-wave properties in the presence of swell. In pure wind waves, there are similarities in the wind-wave growth (Zakharov et al. 2019): the dimensionless wave height and wave period are consistently combined with each other following $H \sim T^R$, in which $H = gH_s/u^2$ and $T = gT_p/u$, where u corresponds to either the wind speed or the frictional speed. Some well-known relations are those proposed by Hasselmann et al. (1976) with $R = 5/3$, Toba (1972) with $R = 3/2$, and Zakharov and Zaslavsky (1983) with $R = 4/3$. Theoretical analyses suggest that these relations are associated with different regimes of wind-wave coupling and represent different stages of wind-wave growth (Badulin and Grigorieva 2012; Badulin and Geogdzhayev 2019), as Hasselmann's relation corresponds to a constant wave momentum flux and represents young waves, while the relation of Zakharov and Zaslavsky indicates the dominance of nonlinear wave-wave interactions corresponding to fully developed waves, and that of Toba corresponds to a constant wind flux into waves, representing an intermediate case between the abovementioned two cases.

Figure 8a shows a plot of the dimensionless wave height against the wave period using the wind-wave information and

wind speed; the relations of Hasselmann et al. (1976) and Zakharov and Zaslavsky (1983) are overlain, Fig. 8b shows a similar dimensionless plot using the friction speed; here, the relation of Toba (1972) is overlain. In Fig. 8a, the best fit to the data points with $H_s/L_p \geq 0.02$ results in an R value of 1.47, closer to that of Zakharov and Zaslavsky (1983). In Fig. 8b, the best fit to the data points with $H_s/L_p \geq 0.02$ results in an R value of 1.50, which almost coincides with that of Toba (1972). The above results indicate that wind waves range from moderately developed to fully developed, which is consistent with the wave age conditions in Fig. 4c. The data points are consistently biased lower than those predicted by the wind-wave growth relations for $H_s/L_p < 0.02$, representing the suppression of wind-wave energy.

There are three main impacts of swell on wind waves. First, swell can interact with the airflow and thus change the wind input (Chen and Belcher 2000), but data points from these cases have been excluded from analysis. The second impact is through the long wave modulation of the wave breaking process (e.g., Dulov et al. 2002); however, this modulation should not result in a reduction in the wind-wave energy because field data implied reduced breaking wave dissipation through whitecap coverage observation in the presence of swell (e.g., Sugihara et al. 2007; Goddijn-Murphy et al. 2011). Third, through the direct interaction between swell and wind wave, field data (Young 2006; Badulin and Grigorieva 2012) and theoretical models (Badulin 2008) have shown strong interactions, as swell can absorb energy from locally generated wind

FIG. 8. (a),(b) Normalized wave height H vs the normalized wave period T . The normalization uses the wind speed in (a) and friction speed in (b). The data are grouped with respect to the wave steepness threshold $H_s/L_p = 0.02$.

wave. The interaction can be related to the dominant wave period or height ratio between swell and wind wave (Masson 1993; Carlsson et al. 2010; Potter 2015).

To analyze the influence of swell on wind-wave growth, a height ratio H_R is defined as the ratio of the measured wind-wave height H_s to that expected from a pure wind wave, which was estimated from the relation of Toba (1972). This ratio was plotted against the frequency ratio ($FR = f_{p_swell}/f_{p_windsea}$) and swell index ($SI = E_{swell}/E_{windsea}$) between swell and wind wave, as shown in Figs. 9a and 9b, respectively. A negative correlation exists between H_R and FR, with correlation coefficients

FIG. 9. Height ratio H_R vs (a) the frequency ratio and (b) the swell index. The error bars represent the standard deviations of the mean.

of -0.38 for the scatter points and -0.95 for the bin-averaged values. No clear trend is observed between H_R and SI, with correlation coefficients of -0.06 for the scatter points and 0.30 for the bin-averaged values. These results indicate that the interaction between wind wave and swell is stronger when their spectral peaks are closer. The height ratio H_R is consistently below the value of 1 for $FR > 0.6$, which is in agreement with the modeling study of Masson (1993). At low FR values ($FR < 0.4$), for which the spectral peaks of swells and wind waves are well separated, the interactions between swells and the energy-containing wind waves are supposed to be rather weak.

5. Discussions and conclusions

The wave state dependence of sea surface roughness and its parameterization have been widely studied for many years, but they are mostly limited to wind-wave-dominant conditions. Directly extrapolating this relationship to mixed waves or swell-dominant conditions would introduce large uncertainty. Considering the frequent occurrence of mixed waves in the open ocean, it is not surprising that a wind-speed-dependent

scaling is still popularly used (e.g., Foreman and Emeis 2010; Andreas et al. 2012; Edson et al. 2013).

In this study, we examined the wave age and wave steepness dependencies of the sea surface roughness using observations taken from a fixed platform. The observations were characterized by moderate wind speeds, moderately developed to fully developed wind waves and mixed wave conditions, which are quite common in the open ocean (e.g., Monahan 2006; Semedo et al. 2011). No obvious trend was found in the wave steepness dependence, using either the full wave spectrum or the wind-wave spectrum alone. The wave age scaling (D03), proposed for pure wind-wave conditions, was found to hold under mixed wave conditions when the wind-wave spectrum was used and the wave steepness was above a threshold ($H_s/L_p > 0.02$), in which the wind-wave properties conform to the wind-wave growth relation. A wind-wave coupling model analysis suggested that the viscous stress tends to decrease with increasing wave steepness due to the airflow separation induced by the breaking of the dominant wind-wave front, which counteracts the increasing trend of the form drag. The H_s/L_p threshold is assumed to be related to the onset of dominant wave breaking.

The development of wind wave is determined by the relative importance of the wind input and wave dissipation, and the nonlinear wave-wave interaction plays a critical role in the self-adjustment process during the evolution of wind waves. As a result, the wind-wave property follows the wind-wave growth relation. In the presence of swell, the measured wind-wave properties follow the wind-wave growth relation quite well for $H_s/L_p > 0.02$, but some deviations are indicated for $H_s/L_p < 0.02$, probably resulting from the nonlinear energy transfer from wind wave to swell. The dependency of the interaction between wind wave and swell on the dominant wave period ratio is illustrated, however, this is a preliminary result, and other possible dependencies on wave characteristic parameters, including the deflection angle and the directional spread and frequency bandwidth of each wave component, should be further diagnosed.

As a result of wave-wave interactions, wind waves are smoothed due to the “lost” energy, reducing the wave-induced stress, but the viscous stress is enhanced due to the reduced airflow separation effect. These combined effects enhance the total drag. Note that the swell effect is quite different from that proposed by García-Nava et al. (2009, 2012a), who suggested that the wind stress is reduced through the impact of swell on short waves. The different wave conditions between their studies and the FOPSCS may account for the different effects of swell. In their studies, the spectral peaks of swells and wind waves are well departed, and thus, the wave-wave interaction is expected to be weak, and the form drag dominates the total wind stress under strong winds and typically young wind waves (Kudryavtsev and Makin 2007; Mueller and Veron 2009). They further showed that the short waves that support the form drag are considerably suppressed by swells. We did not discover the overall attenuation or enhancement of the intensity of short waves in the presence of swells from the FOPSCS (not shown). This may be because swells act to enhance short waves by reducing the AFS effect, which may counteract the suppression of short waves by swell.

The generality of these results should be treated with caution when they were extrapolated to other mixed wave conditions. The observations were characterized by well-developed wind waves at moderate wind speeds, as a result, the wind waves were not as steep as those under strong wind forcing with limited wind fetch or duration (TY01; T12) or those that experience the shoaling effect (Anctil and Donelan 1996), in which the form drag can be dominant (e.g., Suzuki et al. 2014; Babanin et al. 2018). In addition, the swells in the study area are not as old as those propagating thousands of miles across the wide ocean because they are mostly generated within the South China Sea (Li et al. 2018b), and their steepness are enhanced because they experience the shoaling effect near the platform. Therefore, a more comprehensive dataset is required to achieve a better understanding of the wave state dependence of the surface roughness under mixed wave condition.

Acknowledgments. This work is supported by the National Natural Science Foundation of China (41976010, U1706216, 41606024, 41876010, 41806028), the National Key Research and Development Program of China (Grants 2016YFC1402000, 2018YFC1401002), the CAS Strategic Priority Project (XDA19060202), the Public Science and Technology Research Funds Projects of Ocean (201505007), and the Open Fund Project of Key Laboratory of Marine Environmental Information Technology.

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