Special Section:

China-France Oceanography Satellite (CFOSAT): Scientific Applications

Key Points:

- For the first time China France Oceanography Satellite realizes the direct observation of the surface wave-induced stress, and then the wind stress
- The wave-induced stress can increase zonal mean wind stress by more than 30%
- The temporal and spatial variations of wave-induced stress are closely related to the global wind and wave fields

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On the First Observed Wave-Induced Stress Over the Global Ocean

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Abstract Despite many investigations/studies on the surface wave-induced stress, the global feature of the wave-induced stress has not been obtained previously as that requires a simultaneous observation of wave spectra and wind on a global scale. The China France Oceanography Satellite (CFOSAT) provided an opportunity for the first time to evaluate the global wave-induced stress and its contribution to the total wind stress. In this study, the global spatial distributions of wave-induced stress and its correlated index for August to November in 2019 are presented using the simultaneous ocean surface winds and wave spectra from the CFOSAT. The main results show that the wave-induced stress is fundamentally dependent on the wind and wave fields on a global scale and shows significant temporal and spatial variations. Further analyses indicate that there is an upward momentum flux under strong swells and low wind speeds (below ~ 5 m/s), and an anticorrelation between the dimensionless wave-induced stress and the proportion of swell energy to the total. Finally, the variations of the surface wave induced wind stress are clear asymmetric between northern and southern hemispheres in late summer but symmetric in late fall, which are closely associated with the seasonal changes in large-scale atmospheric circulation.

Plain Language Summary Surface waves modulate the wind stress at the air-sea interface, and the modulation processes can be reflected by the wave-induced stress. The global features of wave-induced stress have never been presented due to the extreme difficulty of observing the wave spectra and wind over the world's ocean simultaneously. The China France Oceanography Satellite (CFOSAT) provides an opportunity to investigate this tough issue. This study shows the global monthly mean dimensionless wave-induced stress and the variations of wind stress before and after including surface wave effect. The zonal wave-induced stress can account for 30% of the total wind stress, and it is strongly dependent on the wind and wave states. Additionally, the global patterns of wave-induced stress and wind stress variations with and without wave effect are controlled by the large-scale atmospheric forcing. The results of this study are beneficial to the determination of wind stress parameterization so that the coupled ocean atmosphere models can be improved.

1. Introduction

As one of the most fundamental phenomena at the air-sea interface, surface waves play a significant role in the momentum, heat, and mass fluxes due to its strong interaction with the turbulence of the lower atmosphere and the upper ocean (Babanin et al., 2012; S. Chen et al., 2019; T. S. Hristov et al., 2003; Wu et al., 2017). A comprehensive understanding of surface waves can significantly improve the parameterization of air-sea fluxes to the extent that it can improve the ocean, surface wave, and atmosphere models (Grare et al., 2013; Janssen & Viterbo, 1996; Patton et al., 2019; Wu et al., 2019).

Wind stress (i.e., momentum flux) is a result of Reynolds average turbulence fluctuations, which is modulated by surface waves that overlap with turbulent scales (Sun et al., 2015). Over the ocean, the surface

waves can modulate the atmospheric boundary layer and cause additional velocities $\tilde{u}$, $\tilde{v}$, and $\tilde{w}$ in the airflow, which are defined as wave-coherent or wave-induced velocities (Hristov & Ruiz-Plancarte, 2014). Correspondingly, the total wind stress τ_{tot} above the surface waves should be treated as the sum of the turbulent stress τ_{turb} and the wave-induced stress τ_{wave} , $\tau_{\text{tot}} = \tau_{\text{turb}} + \tau_{\text{wave}}$. Notably, the viscous stress is neglected

because it is only work in a very thin layer near the sea surface (Wu et al., 2017). The wave-induced stress is correlated with the wave age (Hanley & Belcher, 2008). For young wind waves extracting momentum from the local wind, the wind wave-induced stress $\tau_{\text{wave}} > 0$ (Janssen, 1992); however, swells traveling faster than the local wind lose momentum, and then a component with opposite sign is induced $\tau_{\text{wave}} < 0$ (Grachev & Fairall, 2001).

Högström et al. (2015) proposed a simple scheme including the swell-significant wave height (SWH) and the swell peak frequency to calculate the swell-induced stress. Wu et al. (2016) added the swell influence on the wind stress parameterization using this scheme and found that the regional coupled atmosphere-wave model performance is improved concerning the wind speed below 8 m/s. However, the effect is not significant at high wind speeds. Different studies have shown that the calculation of wave-induced stress based on the wave spectra is in line with the physics and practice (S. Chen et al., 2019; Hanley & Belcher, 2008; Janssen, 1992; Semedo et al., 2009; Zou et al., 2018). The wave spectra can be divided into wind-sea and swell parts according to the spectral partitioning method (Högström et al., 2013; Semedo et al., 2011); and then the wind-sea and swell induced stress can be parameterized separately for calculation of the wind stress. S. Chen et al. (2019) showed that the modeled wind stress τ_{tot} is consistent with the observations on two fixed towers according to this scheme.

The global-scale wave climate has received increasing interests in recent decades (G. Chen et al., 2002; Heimbach et al., 1998; Jiang et al., 2019; Semedo et al., 2011; Young, 1999). G. Chen et al. (2002) used combined data sets of satellite scatterometer (SCAT) and altimeter to evaluate the global distribution of the wind-sea- and swell-dominated wave fields and found three tongue-shaped zones of swell dominance in the eastern tropical areas, called “swell pools.” However, they used the wind-wave relation for fully developed seas, not wave spectra. Semedo et al. (2011) used the data set of European Centre for Medium-Range Weather Forecasts (ECMWF) Re-Analysis (ERA-40) including the global and long enough wave spectra information to investigate the global distribution of wind-sea and swell parameters through spectral partitioning. They complemented and validated the results reported by G. Chen et al. (2002). Jiang et al. (2019) investigated the wave climate from spectra of ERA-Interim and concluded the wave connections with local and remote wind fields.

Although the global wind-sea and swell parameters have been determined by the observed and reanalysis data, the global characteristics of wave-induced stress are rarely studied even based on the modeled or re-analysis data. Janssen and Viterbo (1996) discussed the winter season climate of ECMWF model and found the normalized ensemble mean wave-induced stress vary between 0 and 0.8 for the world's ocean. Their results are based on model results and do not consider the swell-induced stress that is negative. Besides, the global distribution of wave-induced stress has not been assessed through the directly observed wave spectrum on a global scale. Only total SWH are provided by altimeter missions, with no spectral information. Although the spectral of waves can be retrieved by synthetic aperture radar (SAR) images, mainly long swell (generally longer than 250 m in wavelength) is accessible (Hauser et al., 2017, 2019). The China France Oceanography Satellite (CFOSAT) observed wave spectrum not only under swell conditions, but also in mixed sea and wind-sea conditions. CFOSAT is able to simultaneously detect the ocean surface winds and wave spectra under different sea state, which provides us a unique opportunity to evaluate the global surface wave-induced stress, and then the total wind stress. As an example of application of the data from CFOSAT, we here investigate the global property of wave-induced stress for the late summer and autumn.

This study is organized as follows: Section 2 describes the CFOSAT, where the longitude and latitude matching method of wind and wave fields and the global ensemble average method are introduced. Results and discussions are presented in Section 3, the patterns and monthly variations of surface waves and wave-induced stress at a global scale are shown, as well as their correlation with large-scale atmospheric processes; the calculation of wave-induced stress is also made in this section. Finally, our concluding remarks are summarized in Section 4.

2. Data and Methods

2.1. CFOSAT

As a new ocean satellite mission, the CFOSAT launched on October 2018 is designed jointly by French and Chinese Space Agencies (CNES and CNSA) to be able to monitor the ocean surface winds and surface

waves over the globe. The CFOSAT embarks two scientific instruments with Ku-band (13.2–13.6 GHz): a wind SCAT to measure wind speed and direction (Zhu et al., 2016); the first real aperture wave SCAT (Surface Wave Investigation and Monitoring instrument, SWIM) to observe the surface waves. Here, the determination of 2-D wave spectrum is the most innovative aspect (Hauser et al., 2017). CFOSAT devotes to the characterization of the ocean surface wind and waves to improve the understanding of the air-sea interaction and to better the forecast of atmosphere, ocean, and wave and their associated applications (Hauser et al., 2019).

The CFOSAT mission chooses a polar orbit, at an altitude of 519 km with a near-real time transmission capability. Since its launch, the calibration of SWIM and SCAT and the validation of data products have been carried out and successfully completed a year later. During fall 2019, the scientific data products are accessible through the AVISO + website. The off-nadir beams 6°, 8°, 10° of SWIM are used to estimate the directional wave spectra and wave parameters in wave cells with the size of ~70 km along track and 90 km across-track. The global coverage of 2-D wave spectra is obtained within 13 days. The wave parameters from beam 10° show a better performance with respect to beams 6° and 8° (Hauser et al., 2017, 2019). Therefore, the wave spectra determined by beam 10° are used in this study. The SCAT operates at larger incidence angles (26°–46° from nadir) and a larger swath to achieve the global wind field with only 3 days, with a spatial resolution of 25 × 25 km or 12.5 × 12.5 km (Zhu et al., 2016). The detailed information about CFOSAT and its products can be found in Hauser et al. (2017, 2019) and AVISO + website, NSOAS website.

2.2. Data Process

So far, the CFOSAT team has published L2 data products of global ocean wind and wave fields for scientific research. The comparisons of SWH H_s and wind speed U with the corresponding parameters from ECMWF model and Jason-3 mission indicated an excellent accuracy for determination of H_s and U . Besides, the 2-D wave spectra from beam 10° showed a good agreement with the performance of ECMWF model and buoys except at very low frequencies (Hauser et al., 2017, 2019). The wave-induced stress is calculated from the wave spectra, here 2-D wave spectra with direction θ from 7.5° to 172.5° by step of 15° and maximum wavenumber $k_{\max}$ of 0.277 m⁻¹ are derived from L2 products. Due to inadequate elimination of noise, a parasite peak with high energy is present at very low frequencies in the wave spectrum (Hauser et al., 2017). Therefore, the wave spectra with the quite low frequencies are eliminated, besides, to capture the swells with a possible maximum period of 25 s (Grachev & Fairall, 2001), the chosen wavenumber range is limited within 0.0065–0.277 m⁻¹.

The data of winds and wave spectra for 4 months from August to November in 2019 are used to analyze the patterns and monthly variations of wave-induced stress and other correlated parameters in this study. Figure 1 shows the track points of central latitudes and longitudes for SWIM wave cells and SCAT wind cells in 1 day over the ocean. The latitudes and longitudes of wind cells can cover the wave cells (Figures 1a and 1b), and we make an exact match between wind and surface wave. Therefore, the latitude and longitude of the wind cell should match with the wave cell, which is the basis for wave spectra partitioning and the estimation of wave-induced stress. First, the time of wave spectra is fixed and the time of the wind field determined within 4 s before and after the wave spectra ($T_{\text{SWIM}} - 4 \text{ s} < T_{\text{SCAT}} < T_{\text{SWIM}} + 4 \text{ s}$) is chosen as the corresponding wind. The second step is to extract the longitude, latitude, and data of the wind field according to the chosen time of SCAT. There are 42 or 84 wind cells in each moment, which depend on the spatial resolution, whereas there is only one wave cell. Therefore, it is necessary to choose the latitude and longitude of the wind cell based on these of the wave cell. The selection criterion is that the sum of the difference between wind and wave cell in longitude plus that in latitude is the minimum. After that, the wind field data corresponds to the wave spectrum data one by one, including time, latitude, and longitude (Figures 1a and 1c).

After the successful match, the 13-day wind speed over the ocean is shown in Figure 2a. Although the 13-day data can cover the global ocean, the overlapping effect is found to make the pattern blurred (Figure 2a), which brings some difficulty when presenting the global variations of the correlated wave parameters clearly. Therefore, we make an ensemble average of the data of waves and winds according to the longitude and latitude grids. Notably, large patterns on a global scale are our focus. Additionally, to avoid the possibility

Figure 1. The track points of central latitude and longitude over the ocean in August 1, 2019. (a) Represents wave cell of SWIM, while (b) and (c) represent wind cell of SCAT before and after matching, respectively.

of data outliers caused by too little data in a single grid, the grid of $3^\circ \times 3^\circ$ is chosen. The rationality of this grid method is illustrated by Figure 2b. Although the data is a monthly average, the large pattern of August in Figure 2b corresponds well to Figure 2a. Besides, we compare the global wind U_{10} distribution in August from CFOSAT with that from ERA5 reanalysis data, and the meridional mean wind speeds (MMU) for both are shown in Figures 2c and 2d. The main patterns for CFOSAT and ERA5 are similar. Some regional wind properties can be well captured such as the Indian Ocean (IO) monsoon and Somalia coastal low-level wind jet, which further demonstrates the similarity between CFOSAT and ERA5. Additionally, the correlation coefficient of MMUs between CFOSAT and ERA5 is ~ 0.83 . All these suggest the feasibility of our method to obtain the monthly mean parameters. Although the correlation between CFOSAT and ERA5 is significant, the difference of the mean MMUs between them is ~ 2.2 m/s with a mean percentage $\sim 30\%$, which is not trivial. In addition, some regional patterns such as South Pacific Ocean are different. This may be due to the limited in situ observation at each point. In view of the above, the global features of wave-induced stress and its correlated parameters will be analyzed and presented based on the synchronously matched monthly mean ocean surface wind and wave data.

3. Results and Discussions

The features of surface wave-induced stress in wind-sea- and swell-dominated wave fields are fundamentally different (Hanley & Belcher, 2008). Therefore, we start with the global mean wavefield characteristics for August to November from the perspective of wind-sea or swell dominance, followed by a detailed description of the surface wave-induced stress and its monthly variations and the difference of wind stresses with and without surface wave effects.

3.1. Swell Index

A normalized index defined by G. Chen et al. (2002) can be used to indicate the wind-sea or swell dominance. The swell index (SI) is the swell-associated energy E_{swell} proportion to the total wave energy E_{tot} at the sea surface:

$$SI = \frac{E_{\text{swell}}}{E_{\text{tot}}} \quad (1)$$

Figure 2. The global wind speed. (a) Represents the 13-day real-time data from August 1st to 13th in 2019 for CFOSAT. (b) and (c) The monthly mean of August from CFOSAT and ERA5 reanalysis data, respectively. (d) The meridional mean wind speed. The blue and red lines correspond to (b) and (c), respectively.

here, E_{swell} can be calculated through the zeroth moments m_0^s of the swell part of wave spectra, $E_{\text{swell}} = \rho_w g m_0^s$, where ρ_w is the water density, g is the gravitational acceleration. The wave energy $E_{\text{tot}} = \rho_w g m_0$, is the sum of swell-associated and wind-sea-associated energy, which indicate that SI can estimate the degree of wind-sea or swell dominance and it can be simplified as m_0^s/m_0 . The 1-D spectrum $E(k)$ is calculated by directional integration of the 2-D spectrum $E(k, \theta)$, and then the swell part is able to be extracted and the zeroth moments m_0^s is calculated according to the method of spectral partitioning provided by Högström et al. (2013).

The globally averaged SI distributions for 4 months are displayed in Figure 3. The spatial patterns vary monthly and geographically and the zonal variations are the striking. A remarkable feature to Figure 3a is that the SI is, to a large extent, coherent with the wind pattern shown in Figure 2b. The agreement of swell variation with the wind field is similar to the result retrieved from SAR wave spectra shown by Heimbach et al. (1998) and the features analyzed from ERA-40 wave spectra shown by Semedo et al. (2011). For the monthly averaged fields, the SI is almost always higher than 0.6 when the wind speeds less than 6 m/s, which indicates a swell dominance. The swells are generally not locally generated, but the wind waves are

Figure 3. The monthly averaged swell index (*SI*) distributions and its zonal variations from (a) August to (d) November in 2019 in the global ocean.

coupled to the local winds so that the wind-sea energy progressively increased with the wind speed, which induces the swell energy proportion to the total wave energy tend to decrease. Additionally, even though only one monthly mean index for the summer in our study, the general pattern is basically the same as the JJA pattern shown in Semedo et al. (2011), but different from the distribution of larger *SI* values mainly concentrated along the western ocean boundary presented in G. Chen et al. (2002). This is possibly induced by the different calculation schemes of swell or wind-sea energy. G. Chen et al. (2002) calculated the swell and wind wave energy according to the relationship between effective height and wave energy, which is an indirect method, not directly based on wave spectrum decomposition.

Figure 4. The variations of the wave-induced stress τ_{wave0} normalized on the turbulent stress at sea surface u_*^2 with (a) the wind speed U_{10} and swell index SI or (b) swell energy normalized by the largest swell energy for the whole month in the global ocean in August, 2019. The black lines in (a) and (b) represent the wind speeds of 5 m/s.

It can be seen from the SI distribution and the zonal average profiles in 4 months that the larger values of SI gradually move from the Northern Hemisphere (NH) to the Southern Hemisphere (SH) with the seasonal change (Figure 3). Close to the equator, the swell energy is mostly greater than 60% of the total energy, which is caused by the lower wind speeds and larger swells in the equatorial climate. The lower values of SI occur at high latitude, in particularly in the SH where the prevailing westerlies dominate during late summer and early fall (Figures 3a and 3b). The physical mechanism behind these features from the global and regional inter-comparisons is that although waves are not purely local, the energy associated with local wind-sea increases with the wind speed; when the local wind decreases, the swells from the distance dominates over the local generated wind waves, which induced a larger SI .

3.2. Surface Wave-Induced Stress

Before the analysis of the features of global mean wave-induced stress, we start with the performance of the wave-induced stress for various wind speeds and swell energies. The calculation of wave-induced stress at the sea surface τ_{wave0} is done in accordance with S. Chen et al. (2019):

$$\tau_{\text{wave0}} = \int_0^{\infty} \frac{C_{\beta} g k u_*^2}{c^2} E(k) dk \quad (2)$$

here, k is the wavenumber; c is wave phase speed; u_* is the friction velocity at the surface; $E(k)$ is the 1-D wave spectrum; C_{β} is the wave growth/decay rate coefficient, $C_{\beta} = 16$ for the wind wave part and $C_{\beta} = -30$ for the swell part following S. Chen et al. (2019). This coefficient is validated by the field observations on fixed tower at two points (S. Chen

et al., 2019). Here, the separation value of the wind-sea and swell spectra parts is defined as the wave phase speed $c = 1.2 U_{10}$ (Pierson & Moskowitz, 1964). The global feature for the open ocean is the main concern. Therefore, for deep water, the cutoff frequency of wave spectra is $f_c = 0.83g / (2\pi U_{10})$, corresponding wavenumber is $k_c = (2\pi f_c)^2 / g$. The wind-sea and swell components can be extracted from the whole spectra above and below this cutoff value, and then the wind-sea and swell-induced stress can be computed by integrating over the respective parts.

The variations of wave-induced stress τ_{wave0} normalized by the turbulent stress at the surface u_*^2 in August are shown in Figure 4. At the wind speed higher than 5 m/s, the overall trend of dimensionless wave-induced stress (DWIS) is increasing with the wind speed U_{10} . DWIS is less than zero at wind speeds below 6–7 m/s, corresponding $SI \sim 0.6$, which identifies the swell-dominated wavefield (Figure 4a). Although at wind speeds below ~ 5 m/s, the swell waves completely dominate the wavefield, the DWIS varies greatly, from -0.2 to approach -2 . The SI reflects the percentage of swell energy in the total, reach to 1 under low wind conditions, which is not able to evaluate the performance of DWIS at low wind speeds. Therefore, the relationship between DWIS and normalized swell energy is presented in Figure 4b. It is clear that the higher the swell energy, the stronger the DWIS if the wind speeds below 5 m/s, which can be identified by Figure 6 in Hanley and Belcher (2008) and Figure 5b in S. Chen et al. (2019). However, the DWIS is more pronounced under relatively larger swell energy when DWIS greater than zero. This may be due to the fact that under larger swell energy, the corresponding wind-sea-associated energy is large enough to produce a considerable positive wave-induced stress.

Given that the observed wave-induced stress shows a strong correlation with the swell energy, we can construct a wave spectrum to see if we can get the consistent or similar behaviors. Therefore, following Hanley and Belcher (2008) and S. Chen et al. (2019), a wave spectrum including wind-sea and swell components is established to theoretically explain the performance. To be consistent with the observational wavenumber

Figure 6. The global distributions of monthly mean dimensionless wave-induced stress from (a) August to (d) November in 2019.

($0.0068 - 0.2770 \text{ m}^{-1}$), the frequency in the constructed wave spectrum is set in the range of $0.0412 - 0.2622 \text{ Hz}$. The swell energy varied with the spectrum peak frequency f_p , the relationship between DWIS and normalized swell energy can be presented naturally by changing f_p (Figure 5). Based on the established wave spectrum, the features of wave-induced stress can be captured, which show excellent agreement with the observations (Figures 4b and 5). Our results show that the performance of DWIS depends on the swell energy in the conditions of swell dominant areas, which can be supported to some extent by the results of Höglström et al. (2015). They proposed a scheme to calculate the swell-induced stress caused by spectra peak parts, which is determined by $H_{sd}^2 f_p^2$, here, H_{sd} is the swell-SWH, which can be calculated by $4\sqrt{m_0^s}$. This also shows that the swell energy is a decisive factor in calculating wave-induced stress.

How to explain the critical wind speed ($\sim 5 \text{ m/s}$) of pure swells and mixed seas is a key question. The wave age c/U_{10} or c/u_* is the decisive parameter for the swell energy proportion to the total energy in a given wave spectrum. Based on the condition for partition of wave spectrum ($c/U_{10} = 1.2$) and the maximum observed wavenumber ($k_{cmax} = 0.2770 \text{ m}^{-1}$) or frequency ($f_{cmax} = 0.2622 \text{ Hz}$) by the SWIM, we can calculate a maximum critical wind speed U_{10cmax} of 4.9373 m/s for deep water through this formula $U_{10cmax} = 0.83g/(2\pi f_{cmax})$. Below this critical wind speeds, the swells fully mask the local wind-seas, and when the swell energy is large enough, the DWIS is less than -1 , which corresponds to an upward momentum flux (Figures 4b and 5). The

DWIS less than -1 occurs at wind speeds of $\sim 5 \text{ m/s}$, which is consistent with the in situ wind stress observations (S. Chen et al., 2018; Drennan et al., 1999). However, the swell energy at low wind speeds is a decisive factor that can be used to interpret the feature of wave-induced stress, which has no any doubt. In the following section, the observed global spectra will be used to calculate the wave-induced stress and to show the global distribution and the influence on wind stress for the first time.

Figure 5. Same as Figure 4b, but for the constructed wave spectra.

3.3. Global Distribution of Wave-Induced Stress

Based on the wave spectra and corresponding wind speeds over the global ocean the DWIS is calculated according to Equation 2, and then the monthly average global distributions of DWIS for August to November are obtained (Figure 6). The global overall patterns of DWIS are generally

Figure 7. The global distributions of monthly mean wind stress for August in 2019. (a) and (b) Represent the data from CFOSAT and ERA5, respectively. (c) The meridional mean wind stress. The blue and red lines correspond to (a) and (b), respectively.

anticorrelated with the performance of SI (Figures 3 and 6), which can also be recognized in Figure 4a. Additionally, the DWIS is strongly coupled to the wind field (Figures 2b and 6a). In the later summer of the NH, the DWIS values vary between -0.9 and 1.1 , where the larger values are found mainly in the IO monsoon zone and the westerlies of the SH; DWIS values less than zero covers most areas of the NH (Figure 6a). The global pattern is changing with the time, the larger values of DWIS are mainly distributed at high latitude of the NH and SH till to the late autumn of the NH with a maximum of 1.2 . Due to the transient variations in sea state are eliminated through monthly average, the DWIS monthly mean values become relatively concentrated. Statistically, in August, DWIS less than zero accounts for more than 50% in the NH, but less than 27% in November. In addition, from August to November, the percentage of DWIS less than zero close to the equator gradually increases from $\sim 56\% - 66\%$, which indicates a swell-dominated wavefield. This swell dominance corresponds to Figure 3, and is similar to the results at low latitudes by Semedo et al. (2011). The physical process behind these global and regional patterns is that the variations of wind and wave fields on a regional to global scale.

After having discussed the global DWIS features, our attention is turned to the differences of the total wind stress with and without surface wave effects at 10 m above the sea surface. Following the steps of S. Chen et al. (2019), we calculate the total stress with and without wave-induced stress and retrieve their difference. We compare the monthly mean wind stress in August derived from CFOSAT with that from the ERA5 reanalysis data, and the global and main regional properties are generally similar to each other except for the South Pacific Ocean (Figures 7a and 7b). The meridional mean wind stress (MMS) in August for CFOSAT and ERA5 are shown in Figure 7c. The correlation coefficient of MMSs between

CFOSAT and ERA5 is as high as 0.85. However, the difference of the mean MMSs is $\sim 0.02 \text{ m}^2\text{s}^{-2}$, and the mean percentage of difference is $\sim 20\%$, which is also not trivial and is similar to the wind speed shown in Figure 2d. Then, the geographical distributions and zonal variations of the percentage of increase or decrease in wind stress are shown in Figure 8 from August to November in 2019. We can see that the increase and decrease of wind stress are mainly concentrated within 40%. However, the increase can be close to or even over 60% in areas covered by strong wind waves; in areas fully masked swells, the wind stress reduces by nearly or even more than 80% relative to that without considering wave impacts (Figure 8d). From the zonal variations, we can see that the significant hemispheric asymmetry is displayed in August; however, the symmetry is impressive in late fall. Again, this is fundamentally caused by the large geographical pattern of wind and wave fields. Notably, the decrease of wind stress after introducing the swell influence is significantly different from the results reported by Janssen and Viterbo (1996). They reported that the normalized wave-induced stress in the global ocean is always greater than zero because the swell-induced stress was not concerned.

The wind stress could be from the ocean to the atmosphere under strong swell conditions (Figures 4 and 5), which has also been observed and simulated in previous studies (S. Chen et al., 2019; Grachev & Fairall, 2001; Nilsson et al., 2012). Additionally, the wind stress will decrease in the areas dominated by swells from a global point of view (Figure 8). Therefore, the parameterization of wind stress is more complicated under swell-dominated conditions than that under wind-sea-dominated conditions. Although the estimation of wind stress can be improved to some extent after introducing the wave influence into roughness length or drag coefficient (Fairall et al., 2003; Guan & Xie, 2004), shift in the direction of upward and downward momentum due to swells is extremely difficult to be directly observed. As stated by Wu et al. (2017), the model simulations can be considerably improved after introducing the swell-induced stress under swell-dominated conditions. Therefore, a feasible way is that the wave influence can be introduced directly into the parameterization of wind stress through the form of wave-induced stress to improve the

Figure 8. The global distributions (left) and zonal variations (right) of the percentage of increase or decrease in wind stress relative to the absence of wave effects for (a) August, (b) September, (c) October, and (d) November in 2019.

model performance, which has been validated by in situ observations and numerical simulations (S. Chen et al., 2019; Semedo et al., 2009).

4. Summary and Conclusions

In this study, the global features of wave-induced stress are for the first time investigated using the simultaneous ocean surface winds and wave spectra observed from the CFOSAT based on the framework of S. Chen et al. (2019). Additionally, the spatial distribution of the wavefield is presented from the perspective of the

swell energy proportion to the total wave energy. The main results indicate that the general features of the wave-induced stress fundamentally depend on the wind and wave fields.

The swell dominance in the overall wavefield varies with the atmospheric forcing and shows a distinct monthly variation. In the late summer, the swell energy even accounts for more than 60% of the total wave energy in the NH, and the larger swell energy mostly covers the low to middle latitude in the fall. It is shown that the wave-induced stress is strongly controlled by the variations of swell energy. At wind speeds below ~ 5 m/s, the upward momentum flux can be generated by strong swells. This critical wind speed is slightly higher than the critical wind speed deduced from in situ observation which is about 4 m/s, which depends on the partitioning schemes of wave spectra and the critical wave age. At wind speeds higher than 5 m/s, the wave-induced stress gradually increases with the wind stronger, in that case wind waves will dominate and the swell energy percentage decline. From the viewpoint of global scale, the monthly spatial patterns for DWIS present an anticorrelation with the SI. A swell dominance is able to be captured by the regional distribution of the wave-induced stress, such as the areas close to the equator. From the zonal variations of the wind stress difference with and without surface wave effect, a hemispheric asymmetry is shown in the late summer, but a significant symmetry in the late fall. The physical picture behind these properties is the seasonal and geographical variations of the wind and wave fields.

Although only the stress characteristics of late summer and autumn are discussed, it clearly shows the significant role of surface waves in the air-sea momentum flux. The wave spectrum partitioning scheme is the key of analyzing the sea state and wave-induced stress. The specific result may change for different wave age criterion (Hanley & Belcher, 2008; Loffredo et al., 2009), but the global pattern is unlikely to change substantially. Although the important role of surface waves in the determination of wind stress can be revealed by using the synchronous wind and wave data derived from the CFOSAT, more field observations including wind stress information are needed to validate in detail to better understand the physical mechanism of wave modulating wind stress.

Data Availability Statement

Data supporting this study are available by any users from <https://www.aviso.altimetry.fr/en/missions/current-missions/cfosat.html>.

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References

- Babanin, A. V., Onorato, M., & Qiao, F. (2012). Surface waves and wave-coupled effects in lower atmosphere and upper ocean. *Journal of Geophysical Research*, 117, C00J01. <https://doi.org/10.1029/2012JC007932>
- Chen, G., Chapron, B., Ezraty, R., & Vandemark, D. (2002). A global view of swell and wind sea climate in the ocean by satellite altimeter and scatterometer. *Journal of Atmospheric and Oceanic Technology*, 19(11), 1849–1859. [https://doi.org/10.1175/1520-0426\(2002\)0192.0.CO;2](https://doi.org/10.1175/1520-0426(2002)0192.0.CO;2)
- Chen, S., Qiao, F., Huang, C. J., & Zhao, B. (2018). Deviation of wind stress from wind direction under low wind conditions. *Journal of Geophysical Research: Oceans*, 123, 9357–9368. <https://doi.org/10.1029/2018JC014137>
- Chen, S., Qiao, F., Jiang, W., Guo, J., & Dai, D. (2019). Impact of surface waves on wind stress under low to moderate wind conditions. *Journal of Physical Oceanography*, 49, 2017–2028. <https://doi.org/10.1175/JPO-D-18-0266.1>
- Drennan, W. M., Kahma, K. K., & Donelan, M. A. (1999). On momentum flux and velocity spectra over waves. *Boundary-Layer Meteorology*, 92(3), 489–515. <https://doi.org/10.1023/a:1002054820455>
- Fairall, C. W., Bradley, E. F., Hare, J. E., Grachev, A. A., & Edson, J. B. (2003). Bulk parameterization of air-sea fluxes: Updates and verifications of the COARE algorithm. *Journal of Climate*, 16(4), 571–591. [https://doi.org/10.1175/1520-0442\(2003\)016<0571:BPOASF.2.0.CO;2](https://doi.org/10.1175/1520-0442(2003)016<0571:BPOASF.2.0.CO;2)
- Grachev, A. A., & Fairall, C. W. (2001). Upward momentum transfer in the marine boundary layer. *Journal of Physical Oceanography*, 31(7), 1698–1711. [https://doi.org/10.1175/1520-0485\(2001\)031<1698:umtftm>2.0.co;2](https://doi.org/10.1175/1520-0485(2001)031<1698:umtftm>2.0.co;2)
- Grare, L., Lenain, L., & Melville, W. K. (2013). Wave-coherent airflow and critical layers over ocean waves. *Journal of Physical Oceanography*, 43, 2156–2172. <https://doi.org/10.1175/JPO-D-13-056.1>
- Guan, C., & Xie, L. (2004). On the linear parameterization of drag coefficient over sea surface. *Journal of Physical Oceanography*, 34(12), 2847–2851. <https://doi.org/10.1175/jpo2664.1>
- Hanley, K. E., & Belcher, S. E. (2008). Wave-driven wind jets in the marine atmospheric boundary layer. *Journal of the Atmospheric Sciences*, 65(8), 2646–2660. <https://doi.org/10.1175/2007jas2562.1>
- Hauser, D., Tison, C., Amiot, T., Delaye, L., Corcoran, N., & Castellan, P. (2017). SWIM: The first spaceborne wave scatterometer. *IEEE Transactions on Geoscience and Remote Sensing*, 55(5), 3000–3014. <https://doi.org/10.1109/TGRS.2017.2658672>
- Hauser, D., Tourain, C., & Lachiver, J. M. (2019). CFOSAT: A new mission in orbit to observe simultaneously wind and waves at the ocean surface. *Space Research Today*, 206, 15–21. <https://doi.org/10.1016/j.srt.2019.11.012>
- Heimbach, P., Hasselmann, S., & Hasselmann, K. (1998). Statistical analysis and intercomparison of WAM model data with global ERS-1 SAR wave mode spectral retrievals over 3 years. *Journal of Geophysical Research*, 103(C4), 7931–7977. <https://doi.org/10.1029/97JC03203>

- Högström, U., Rutgersson, A., Sahlée, E., Smedman, A. S., Hristov, T. S., Drennan, W. M., & Kahma, K. K. (2013). Air-sea interaction features in the Baltic sea and at a Pacific trade-wind site: An inter-comparison study. *Boundary-Layer Meteorology*, 147(1), 139–163. <https://doi.org/10.1007/s10546-012-9776-8>
- Högström, U., Sahlée, E., Smedman, A. S., Rutgersson, A., Nilsson, E., Kahma, K. K., & Drennan, W. M. (2015). Surface stress over the ocean in swell-dominated conditions during moderate winds. *Journal of the Atmospheric Sciences*, 72(12), 4777–4795. <https://doi.org/10.1175/JAS-D-15-0139.1>
- Hristov, T. S., Miller, S. D., & Friehe, C. A. (2003). Dynamical coupling of wind and ocean waves through wave-induced air flow. *Nature*, 422(6927), 55–58. <https://doi.org/10.1038/nature01382>
- Hristov, T., & Ruiz-Plancarte, J. (2014). Dynamic balances in a wavy boundary layer. *Journal of Physical Oceanography*, 44(12), 3185–3194. <https://doi.org/10.1175/JPO-D-13-0209.1>
- Janssen, P. A. E. M. (1992). Experimental evidence of the effect of surface waves on the airflow. *Journal of Physical Oceanography*, 22(12), 1600–1604. [https://doi.org/10.1175/1520-0485\(1992\)0222.0.CO;2](https://doi.org/10.1175/1520-0485(1992)0222.0.CO;2)
- Janssen, P. A. E. M., & Viterbo, P. (1996). Ocean waves and the atmospheric climate. *Journal of Climate*, 9(6), 1269–1287. [https://doi.org/10.1175/1520-0442\(1996\)009<1269:OWATAC>2.0.CO;2](https://doi.org/10.1175/1520-0442(1996)009<1269:OWATAC>2.0.CO;2)
- Jiang, H., & Mu, L. (2019). Wave climate from spectra and its connections with local and remote wind climate. *Journal of Physical Oceanography*, 49(2), 543–559. <https://doi.org/10.1175/JPO-D-18-0149.1>
- Loffredo, L., Monbaliu, J., Bitner-Gregersen, E., & Toffoli, A. (2009). The role of spectral multimodality in wave climate design. 11th International Workshop on Wave Hindcasting and Forecasting and Coastal Hazard Symposium. Halifax, NS: Canada, p. 10.
- Nilsson, E. O., Rutgersson, A., Smedman, A. S., & Sullivan, P. (2012). Convective boundary-layer structure in the presence of wind-following swell. *Quarterly Journal of the Royal Meteorological Society*, 138(667), 1476–1489. <https://doi.org/10.1002/qj.1898>
- Patton, E. G., Sullivan, P. P., Kosovic, B., DudhiaMahrt, J. L., Zagar, M., & Maric, T. (2019). On the influence of swell propagation angle on surface drag. *Journal of Applied Meteorology and Climatology*, 58(5), 1039–1059. <https://doi.org/10.1175/JAMC-D-18-0211.1>
- Pierson, W. J., & Moskowitz, L. (1964). A proposed spectral form for fully developed wind seas based on the similarity theory of S. A. Kitai-gorodskii. *Journal of Geophysical Research*, 69(24), 5181–5190. <https://doi.org/10.1029/JZ069i024p05181>
- Semedo, A., Saetra, Ø., Rutgersson, A., Kahma, K., & Pettersson, H. (2009). Wave-induced wind in the marine boundary layer. *Journal of the Atmospheric Sciences*, 66(8), 2256–2271. <https://doi.org/10.1175/2009JAS3018.1>
- Semedo, A., Sušelj, K., Rutgersson, A., & Sterl, A. (2011). A global view on the wind sea and swell climate and variability from era-40. *Journal of Climate*, 24(5), 1461–1479. <https://doi.org/10.1175/2010JCLI3718.1>
- Sun, J., Nappo, C. J., Mahrt, L., Belušić, D., Grisogono, B., Stauffer, D. R., et al. (2015). Review of wave-turbulence interactions in the stable atmospheric boundary layer. *Reviews of Geophysics*, 53(3), 956–993. <https://doi.org/10.1002/2015RG000487>
- Wu, L., Breivik, O., & Rutgersson, A. (2019). Ocean-wave-atmosphere interaction processes in a fully coupled modelling system. *Journal of Advances in Modeling Earth Systems*, 11(11), 3852–3874. <https://doi.org/10.1029/2019MS001761>
- Wu, L., Rutgersson, A., & Nilsson, E. (2017). Atmospheric boundary layer Turbulence closure scheme for wind-following swell conditions. *Journal of the Atmospheric Sciences*, 74(7), 2363–2381. <https://doi.org/10.1175/JAS-D-16-0308.1>
- Wu, L., Rutgersson, A., Sahlée, E., & Guo Larsén, X. (2016). Swell impact on wind stress and atmospheric mixing in a regional coupled atmosphere-wave model. *Journal of Geophysical Research: Oceans*, 121(7), 4633–4648. <https://doi.org/10.1002/2015JC011576>
- Young, I. R. (1999). Seasonal variability of the global ocean wind and wave climate. *International Journal of Climatology*, 19(9), 931–950. [https://doi.org/10.1002/\(SICI\)1097-0088\(199907\)19:9<931::AID-JOC412>3.0.CO;2-O](https://doi.org/10.1002/(SICI)1097-0088(199907)19:9<931::AID-JOC412>3.0.CO;2-O)
- Zhu, D., Dong, X., Yun, R., & Xu, X. (2016). Recent advances in developing the CFOSAT scatterometer. IEEE International Geoscience and Remote Sensing Symposium (IGARSS), Beijing, China (pp. 5801–5803).
- Zou, Z., Zhao, D., Zhang, J., Li, S., Cheng, Y., Lv, H., & Ma, X. (2018). The influence of swell on the atmospheric boundary layer under nonneutral conditions. *Journal of Physical Oceanography*, 48(4), 925–936. <https://doi.org/10.1175/JPO-D-17-0195.1>