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Can contemporary satellites estimate swell dissipation rate?

Haoyu Jiang[a,b,*], Alexander V. Babanin[c], Qingxiang Liu[c], Justin E. Stopa[d], Bertrand Chapron[d], Ge Chen[e,b]

[a] College of Marine Science and Technology, China University of Geosciences, Wuhan, China
[b] Laboratory for Regional Oceanography and Numerical Modeling, Qingdao National Laboratory for Marine Science and Technology, Qingdao, China
[c] Melbourne School of Engineering, University of Melbourne, Melbourne, Australia
[d] Laboratoire d'Océanographie Physique et Spatiale, Centre de Brest, IFREMER, Plouzané, France
[e] Department of Marine Technology, College of Information Science and Engineering, Ocean University of China, Qingdao, China

A R T I C L E I N F O

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A B S T R A C T

Swell dissipation can influence air-sea interactions and is an error source in numerical models. Remote sensors, including altimeters and synthetic aperture radars, were employed to estimate swell dissipation rates in previous studies. A detailed error analysis is conducted here to better understand the results of these studies. With the help of a numerical model, we find that the point source model cannot separate swell dissipation from frequency dispersion and angular spreading effectively. Monte Carlo simulations of remotely sensing an ideal swell packet show that the accuracy of the estimated dissipation rate depends on the number, the span, and the wave height error of the observations on the swell track. The results also demonstrate that the dependence between estimated dissipation rates and wave steepness is an inherent relation from the fitting process which cannot prove that the boundary-layer turbulence is the leading reason for swell dissipation. Both the point source and the measurement errors of wave heights cause large errors of estimated dissipation rates. Comparing the overall error with the estimated dissipation rates, this study shows that the present swell tracking scheme using satellite data is sufficient to estimate a mean dissipation rate statistically through many swell tracks but is insufficient for case studies of swell dissipation. Besides, a buffer analysis shows that islands might cause more potential errors in many swell tracks used in previous studies, and ideal locations for future experiments are also identified. More advanced techniques are required for a better understanding of swell dissipation.

1. Introduction

Strong storms in the ocean can generate long gravity waves that propagate out of their sources. After propagating away from their generation areas, these waves are not or weakly affected by the local wind and are known as swells. Swells propagate with a nearly constant period and group speed along great circles of the Earth surface, closely following the dispersion relation and principles of geometrical optics in deep water without currents (Snodgrass et al., 1966; Collard et al., 2009). Long swells can propagate over large distances, radiating momentum and energy across ocean basins (Munk et al., 1963). Therefore, as a ubiquitous phenomenon in the open ocean, swells are observed to dominate the wave condition on > 75% of the time and energy over the globe (Chen et al., 2002; Jiang and Chen, 2013).

Beginning with the needs of forecasting waves in wars, studies of swells have been conducted for more than a century and are still of interest because the uncertainty of swells is still an error source in wave models (WISE Group, 2007). An important topic about the ocean swell is its attenuation and dissipation. Given the fact that swells generated in the Southern Ocean can propagate across the whole Pacific before reaching the coast of Alaska (Snodgrass et al., 1966), swell dissipation in deep water should be weak comparing with the dissipation due to bottom friction and wave breaking of wind-seas (Ardhuin et al., 2009). In spite of this, dissipation is considered as an important process during swell propagation, because it is potentially related to air-sea interactions (e.g., Högström et al., 2013), ocean mixing (e.g., Babanin, 2006), and wave-current-turbulence interactions (e.g., Babanin, 2012) and it is an important source term for numerical wave models (e.g., Zieger et al., 2015).

There are only a few quantitative studies about swell evolution over large distances, thus, swell dissipation is still poorly understood. Laboratory studies and numerical experiments provide some plausible mechanisms for swell dissipation (e.g. Babanin and Haus, 2009; Perignon et al., 2014). Some studies of air-sea interactions show that

* Corresponding author at: College of Marine Science and Technology, China University of Geosciences, 388 Lumo Road, Wuhan, China.
E-mail address: HaoyuJiang@cug.edu.cn (H. Jiang).

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swells can transfer momentum into the lower atmosphere (e.g., Kudryavtsev and Makin, 2004; Semedo et al., 2009; Högström et al., 2013). But studies based on observations are indispensable for investigating such a transoceanic phenomenon. The first systematic field experiment on swell evolution is conducted by some pioneers of modern physical oceanography. In the landmark work of Snodgrass et al. (1966), a sensor array was arranged along a great circle in the Pacific Ocean to trace the swell evolution and to estimate the swell dissipation rate. From today's perspective, their method faces two main shortcomings: few swells exactly line up with the measurement array and the island sheltering introduces large errors. However, little progress was made on their results in the next few decades (WISE Group, 2007).

Ocean remote sensing brings some new opportunities to this issue because satellites can provide measurements with a quasi-global coverage. After the launch of Seasat, remote sensing data begin to play an increasingly important role in ocean observation. Space-borne observations from altimeters and Synthetic Aperture Radars (SARs) are proved to be useful in studies of swells (e.g., Chen et al., 2002; Heimbach et al., 1998). Holt et al. (1998) and Heimbach and Hasselmann (2000) demonstrate that the SAR can track the swells across the ocean using the dispersion relation along the great circle. Based on this, another breakthrough about swell dissipation is conducted by Collard et al. (2009) and Ardhuin et al. (2009) using ENVISAT Advanced SAR (ASAR) swell significant wave heights (SWHs) (Chapron et al., 2001). They give an estimation of swell dissipation rate based on 22 swell events and yield significantly better outputs by applying their results in numerical wave models (Ardhuin et al., 2010; Stopa et al., 2015). After that, studies of estimating swell dissipation rates are also presented using other data from altimeters or SARs (e.g., Young et al., 2013; Jiang et al., 2016; Stopa et al., 2016).

Using data of remote sensing to estimate swell dissipation rates has many potential errors. Although errors are unavoidable in any quantitative physical experiments, a noteworthy feature of the aforementioned errors is that the dissipation rate itself is small so that the results of these observational studies may be greatly influenced by these errors (Jiang et al., 2016). A systematic and quantitative error analysis of estimating swell dissipation using satellites which is the aim of the present paper will be helpful to better explain observed results, to better understand swell dissipation, and to guide future experiments.

2. Swell evolution and point source

The swell energy distribution can be described as wave spectral densities $G(t, \varphi, \lambda, f, \theta)$ with given time t, latitude φ, longitude λ, frequency f and direction θ. Following the dispersion relation, the spectral density propagates along great circles on the Earth surface with a group speed of $C_g = g / 4\pi f$ in deep water without currents (Barber and Ursell, 1948). After normalizing, the spatial evolution rate of spectral density can be expressed by Lagrangian method:

$$I_d = -\frac{\mathrm{d}G(t,\varphi,\lambda,f,\theta)/\mathrm{d}x}{G(t,\varphi,\lambda,f,\theta)} \tag{1}$$

where the x is the propagation distance and the negative sign is to make the rate positive when the energy is decreasing. For long swells at a few storm radii outside the sources without strong winds along the track, wave-wave interactions are negligible (Hasselmann, 1963). When there is no wave breaking and the swell track is far from shallow waters or islands, this spatial evolution rate is equivalent to the non-breaking swell dissipation rate (hereafter, the term "swell dissipation" refers to non-breaking swell dissipation). The above hypotheses, including the absence of currents, deep waters, being away from islands, and negligible wave-wave interactions, are taken for granted for the rest of the analysis.

For satellites, the spectral resolution of contemporary SARs' wave mode (ERS-1/2 and ENVISAT) is insufficient for estimating swell dissipation rates using the spectral distribution (Collard et al., 2009). Altimeters can only retrieve the total SWH, but this SWH can be regarded as the swell SWH if there is only one swell partition in the nadir without wind-seas (Jiang et al., 2016). The swell energy E_s is still the only parameter employed to estimate swell dissipation from the remote sensing platforms:

$$E_s(t,\varphi,\lambda) = \int_0^{2\pi}\int_0^{\infty} G_s(t,\varphi,\lambda,f,\theta)\,\mathrm{d}f\,\mathrm{d}\theta \tag{2}$$

where G_s is the spectral densities of the swell system being integrated. For spectra with more than one partitions, spectral partitioning (e.g., Hanson and Phillips, 2001) is needed before integrating.

Similar to Eq. (1), the spatial energy evolution rate of a swell packet can be defined as $-\frac{1}{E_s}\frac{\mathrm{d}E_s}{\mathrm{d}x}$ following the group speed of $C_g = g / 4\pi f_p$ where f_p is the peak frequency, as the propagation of swell packets also closely follow the dispersion relation along geodesics. In spite of the similar definitions, the spatial energy evolution rate is different from the swell dissipation rate. For ocean swells, the total attenuation of wave energy along the propagation path can be regarded as the superposition of the effects of frequency dispersion, angular spreading, and energy dissipation:

$$-\frac{1}{E_s}\frac{\mathrm{d}E_s}{\mathrm{d}x} = I_{fd} + I_{as} + I_d \tag{3}$$

where I_{fd} and I_{as} correspond to the energy attenuation of the swell packet due to frequency dispersion and angular spreading, respectively. The effects of the three terms on the right-hand side of Eq. (3) are illustrated in Fig. 1. After propagating over a distance of dx, frequency dispersion (Fig. 1a) and angular spreading (Fig. 1b) reduce the energy of a swell packet by narrowing the spectrum in corresponding axes, but the spectral density at f_p and peak direction θ_p stays the same in these two processes. Meanwhile, swell dissipation reduces the spectral densities in all frequencies (Fig. 1c) and directions (Fig. 1d) without narrowing the spectrum.

It is nearly impossible to give the analytical expressions of I_{fd} and I_{as} because they depend on the swell source characteristics, differing from storm to storm. In numerical models, the error caused by the discretization of frequencies and directions during propagation is called the "garden sprinkler effect". Numerical methods are presented to correct this effect (e.g., Tolman, 2002), which is tantamount to the

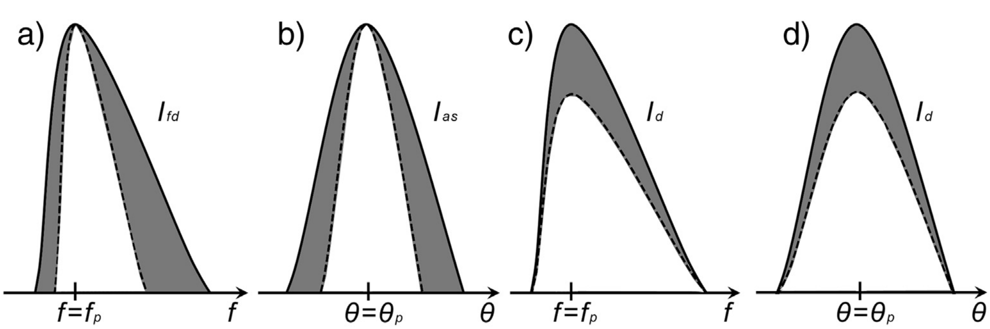

Fig. 1. The variations of the wave spectrum of a swell packet after propagating over a small distance of dx corresponding to (a) frequency dispersion (shown in the frequency spectrum), (b) angular spreading (shown in the directional spectrum), and (c) (d) swell dissipation (shown in both the frequency spectrum and the directional spectrum). The spectra before and after propagating are shown in solid and dash lines, respectively. It is noted that all the shadowed regions showing the energy decrements are exaggerated.