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The Reynolds wave shear stress in partially reflected waves

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ABSTRACT

Wind–waves play a relevant role in the downward flux of mass, momentum and heat, and in their balances. Thus, they are quite significant in small- and large-scale processes like global warming, especially in continental shelf. There is still no consensus on the effectiveness of different mechanisms in enhancing, e.g., the carbon dioxide (CO_2) uptake in the continental shelves with respect to the open ocean. To our best knowledge, the role of partial reflections of short and long waves in the budget has been largely neglected without a specific justification, even though reflection in water waves is a recurring phenomenon. Presumably, this is a consequence of the difficulties in performing experiments with variable reflection coefficient and phase shift of the waves approaching coastal structures. We report experiments on the vertical momentum transfer in partially reflecting waves propagating on a flat horizontal bottom. We measured the free surface level and the fluid velocity, and estimated the Reynolds wave–generated shear stress. A set of experiments using paddle waves or a paddle plus wind waves and a reflection coefficient ranging from $0.10 - 0.75$, with a phase shift of $\pi/4 - \pi$ rad, confirms the theoretical model. The results indicate that reflection must be carefully considered in the correct interpretation of data, particularly when considering momentum transfer in the vertical. The new contributions in the paper are (i) the adoption of a generation–absorption system that controls the reflection coefficient and phase shift between incident and reflected waves; (ii) the quantification of the phase shift influence on the vertical momentum transfer and its horizontal variability; (iii) the consequences for locally–generated wind–waves and wind generated surface currents. To our knowledge, these are the first experiments where the characteristics of the reflected waves have been imposed, allowing a thoroughly analysis of their effects on the flow field.

1. Introduction

Reflection and partial reflection of waves, wave groups, infra gravity waves, long waves are widespread and diffused in coastal regions and, more broadly, in continental shelf areas. In fact, in the continental shelf sediment transport, bed forms morphology and chemical exchanges, both in the sea–bed interface and in the air–sea interface to and from the atmosphere, play major roles in the overall scenario.

The role of continental shelves in the global cycles of several chemicals, like carbon dioxide (CO_2), is widely recognised as important and is subject to numerous analyses e.g., Fennel (2010). The carbon cycle has been influenced by human activities and from changes of the fluxes at the air–sea interface, in particular near the coasts (Mackenzie et al., 2011). The quantification of the changes is generally subject to a large uncertainty (50–100%) (Bauer et al., 2013), due to the numerous phenomena influencing the balance and to the reduced accuracy of field measurements, in particular in the Arctic shelves areas. The regional studies on CO_2 balance are highly dispersed, due, e.g., to non-homogeneity in the definitions of continental shelf and to different methodologies used in these studies. However, even taking into account the relevant data dispersion, bulk evaluations indicate that the average flux densities in the continental shelf area (-0.7 to -1.2 mol C m^{-2} yr^{-1}) are higher than in the open ocean (-0.5 mol C m^{-2} yr^{-1}) (Laruelle et al., 2014). This trend has been confirmed by more sophisticated approaches, where the balance of CO_2 is obtained by including direct surface ocean CO_2 measurements, atmospheric CO_2, in a model with gas transfer inferred from wind speed, salinity and temperature. These last recent models (Laruelle et al., 2014) confirm that sink of CO_2 in the continental shelf areas is larger than in open ocean (up to 40%, with an estimated uncertainty of less than 30%).

In general, there is still no consensus about many aspects involved in continental shelf areas budget of CO_2. The adoption of a global model is limited by the low grid resolution for simulating many processes, since coarse-resolution global models fail in reproducing coastal-ocean circu-

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lation (Fiechter et al., 2014), upwelling, coastal currents, which affect the variability of air–sea CO_2 fluxes in these areas (Borges, 2005; Lachkar et al., 2007). Notice that reflection, in natural environments, is always partial reflection and often takes place in the coastal areas and generally in the continental shelf areas for long waves. We infer that partially–reflected waves play a role in the circulation and locally affect carbon dioxide (and other chemicals) exchanges.

Reflection also contributes to the structural change of the bottom boundary layer in shallow water, with consequent effects on sediment transport and on bed forms. Partial reflection introduces a further scale in a flow field already controlled by several effects (Sánchez-Badorrey et al., 2009), and may cause toe erosion and the generation of rhythmic bed forms (Sánchez-Badorrey et al., 2008). The interaction between incoming and reflected wave trains can modify the sediment transport mechanism. This event eventually enhances scouring and favors the failure of break waters and protective structures (Baquerizo and Losada, 1999a). In the real world, breakwaters and sea-walls always generate wave reflection (rubble-mound breakwaters generally reflect approximately 30–40%, vertical break-waters always higher than 70%), thus partial reflection and its effects are almost ubiquitous. Non-linear waves modulated by partial reflection, and currents (always present in the field) change the intensity of sediment transport (Ribberink et al., 1995).

In the past decades numerous researchers have performed the analysis of vertical momentum transfer in sea gravity waves in different conditions, with waves subject to growth due to the wind action, to shoaling or to decay. This transfer is quite important to describe the flow field and to quantify all the exchanges inside the water column and at the interface with the air. Wave growth mechanism in the presence of wind has been long debated (Phillips, 1957; Miles, 1957; Longuet-Higgins, 1969). Whatever is the action of the wind (tangential stress, fluctuating pressure and resonance phenomena), a net transfer of momentum from the air side toward the water is expected.

The general description of the flow field takes advantage of the triple decomposition (Phillips, 1966) in the sense of Thais and Magnaudet (1995). The variables of interest are split in a time–averaged, a periodic (organized) and a fluctuating (turbulent) component, e.g., $u = \overline{u} + \tilde{u} + u'$ for the horizontal velocity, and $w = \overline{w} + \tilde{w} + w'$ for the vertical velocity. The fluctuating component is calculated as the difference between the raw signal and the sum of the time average and of the phase average terms. By combining these expressions in the horizontal momentum equation and time averaging, it results that the shear stress in the vertical is the combination of several covariances between the velocity components. Among them, we are specifically interested in the term $-\overline{\tilde{u}\tilde{w}}$, herein defined Reynolds wave(–generated) shear stress, which in the present experiments is more than one order of magnitude larger than the turbulent shear stress $-\overline{u'w'}$. In progressive waves advancing on a flat horizontal bottom without source or sink of energy, the term $-\overline{\tilde{u}\tilde{w}}$ is null, since $\tilde{u}$ and $\tilde{w}$ are in quadrature. In dissipative water waves (Deigaard and Fredsøe, 1989) the change of wave conditions in the direction of propagation gives an out–of–quadrature contribution to $-\overline{\tilde{u}\tilde{w}}$, which is used to explain the fluxes of energy and the efficiency of vertical momentum exchange. In particular, the vertical profile of the Reynolds wave shear stress is linear for dissipation near the surface and near the bottom. The sloping bottom also acts inducing a linear profile of the Reynolds wave shear stress (De Vriend and Kitou, 1991). Rivero and Arcilla (1995) analyzed the contribution of various sources to $-\overline{\tilde{u}\tilde{w}}$ and gave an interpretation of this term by introducing the oscillatory vorticity $\overline{\tilde{w}\tilde{\omega}}$ ($\tilde{\omega}$ is orthogonal to the plane of u and w). Their results collapse to the results of other researchers for irrotational flows. See also the analysis in De Serio and Mossa (2013) for irregular shoaling waves.

A recent paper by Olfateh et al. (2017) is focused on Reynolds wave–generated shear stress in wind–generated waves. The Authors describe some contradicting experimental results in literature, showing variegate data on the contribution of the Reynolds wave shear stress to the momentum transfer under wind generated waves. They suggest the reflection effects as responsible of varying sign and magnitude of the experimental values of $-\overline{\tilde{u}\tilde{w}}$, in addition to possible contamination due to recirculation cells in the experimental channels. However, they do not consider the phase and their results are horizontally averaged.

In this research we analyze the Reynolds wave shear stress in a flow field where a partial reflection takes place. We separate the role of the module of the reflection coefficient K_r, of the phase shift and the contribution, if any, of the local generated wind wave (period, length and height, and if is breaking or not) and of the water surface current. This flow field is quite diffuse in natures, since porous structures, beaches, uneven bottom, are all sources of reflected waves, but it is quite complex to reproduce in the laboratory since the variation of the reflection coefficient requires a change of the end wall characteristics of the flume, and the control of the phase shift is almost impossible to realize. We take advantage of a two–paddle flume recently installed in Granada, where the paddle control allows a variation of the reflection coefficient from zero to unity, also in presence of wind and with current recirculation, offering numerous combinations of flow conditions. Here we analyze regular waves without or with wind action. The extension to random waves and to waves plus currents is left for future work.

We remark the importance of our laboratory data, since the same type of information is hard to be obtained in the field, due e.g. the complex pattern of natural waves or to the structure of the reflecting elements.

The paper is structured as follows. In Section 2 the theoretical problem is set, evaluating the covariance $-\overline{\tilde{u}\tilde{w}}$ and the radiation stress component useful for estimating the mean water level profile. In Section 3 the experimental set-up and the procedure for executing the experiments are illustrated, while in Section 4 the measured covariances and mean water levels are compared with theory. The conclusions are reported in Section 5.

2. Theory

2.1. Wave induced Reynolds stress for regular reflected waves in uniform depth

In this section we consider only the (potential) wave-induced terms mainly following the approach by Goda and Abe, 1968. The potentials of intersecting waves can be expressed by the sum of the independent potentials for linear theory. However, non-linear theory requires a more in–depth analysis to include the interaction terms, often described as "bound" waves. The potential and the free surface elevation of two trains of finite amplitude waves propagating in the opposite direction are

$$\Phi = \Phi_i + \Phi_r + \Phi_b, \quad \eta = \eta_i + \eta_r + \eta_b, \tag{1}$$

where i, r and b stand for "incident", "reflected" and "bound", respectively. The potential and the free surface elevation are expanded in series as

$$\Phi_i = a\Phi_i^{(0)} + a^2\Phi_i^{(1)}, \quad \Phi_r = K_r a\Phi_r^{(0)} + K_r^2 a^2\Phi_r^{(1)}, \quad \Phi_b = a^2\Phi_b^{(1)}, \tag{2}$$

$$\eta_i = a\eta_i^{(0)} + a^2\eta_i^{(1)}, \quad \eta_r = K_r a\eta_r^{(0)} + K_r^2 a^2\eta_r^{(1)}, \quad \eta_b = a^2\eta_b^{(1)}, \tag{3}$$

where K_r is the ratio of the reflected wave amplitude to the incident wave amplitude, and a is the incident wave amplitude. The incident and the reflected waves satisfy the classical differential problem

$$\left.\begin{array}{l}
\Phi_{,xx} + \Phi_{,zz} = 0, \quad -h < z < \eta, \quad \forall x, \\
g\eta + \Phi_{,t} + \dfrac{1}{2}\left[(\Phi_{,x})^2 + (\Phi_{,z})^2\right] \quad \text{on } z = \eta, \\
\Phi_{,z} = \eta_{,t} + \eta_{,x}\Phi_{,x} \qquad \text{on } z = \eta, \\
\Phi_{,z} = 0 \quad \text{on } z = -h,
\end{array}\right\} \tag{4}$$

with the additional conditions