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Boussinesq modeling of spatial variability of infragravity waves on fringing reefs

Shih-Feng Su [a,*], Gangfeng Ma [b], Tai-Wen Hsu [c]

[a] Department of Water Resources and Environmental Engineering, Tamkang University, Taipei, Taiwan
[b] Department of Civil and Environmental Engineering, Old Dominion University, Norfolk, VA, USA
[c] Department of Harbor and River Engineering, Research Center for Ocean Energy and Strategies, National Taiwan Ocean University, Keelung, Taiwan

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ABSTRACT

Spatial variations of infragravity waves on fringing reefs are studied using a fully nonlinear Boussinesq equation model FUNWAVE-TVD. The effects of bottom roughness, forereef slope and tidal water level on significant wave height, wave setup and infragravity wave motion are investigated. Model results show that the cross-shore distributions of significant wave heights and setups are reasonably reproduced using the calibrated bottom friction coefficient. However, the infragravity wave heights over the reef flat are under-predicted. Spatial variations of infragravity waves at different tidal water levels exhibit distinct patterns. Spectral peaks and valleys on the reef flat, indicative of the generation of standing infragravity waves, are observed in the measurements and reproduced by the model. It is demonstrated that the model is capable of simulating the generation and propagation of infragravity motions on fringing reefs. The effects of relative submergence and inverse wave steepness on infragravity waves are investigated through a series of numerical experiments. The results demonstrate that the relative submergence has significant effects on the distributions of infragravity waves on the fringing reef. The maximum infragravity wave height decreases with increasing relative submergence at the reef edge. However, it increases with increasing relative submergence at the inner reef flat.

1. Introduction

Waves with infragravity periods (typically 20–200 s), relatively longer than the periods of short swell and sea waves (roughly 4–20 s), have been not only observed on sloping sandy beaches (Bowen and Guza, 1978; Elgar et al., 1992; Guza and Thornton, 1985; Herbers et al., 1994, 1995; Holman, 1981; Huntley et al., 1981; Okihiro et al., 1992), but also in coral reef environments, which are characterized by abrupt bathymetric change in the forereef slope and rough bottom at the reef flat. Field observations have shown that infragravity waves become the dominant wave components on the reef flat, where swell and sea waves have lost most of their energies due to depth-limited wave-breaking and bottom frictional dissipation (Brander et al., 2004; Collen et al., 2009; Hardy and Young, 1996; Lugo-Fernández et al., 1998; Samosorn and Woodroffe, 2008). Infragravity waves play a critical role in nearshore circulations (Kobayashi and Karjadi, 1996), nutrient uptake, sediment transport (Aagaard and Greenwood, 1994, 2008; Beach and Sternberg, 1988; Roberts and Suhayda, 1983) as well as morphologic evolution (Aagaard, 1990; Bauer and Greenwood, 1990; Rockliff and Smith, 1985). In addition, the long-period infragravity motions may produce resonant oscillations on the reef flat to increase water level at the shoreline (Péquignet et al., 2009) and thus enhance beach swash (Holman et al., 1978).

The generation mechanisms of infragravity waves have been well studied by researchers. One of the fundamental mechanisms was proposed by Longuet-Higgins and Stewart (1962, 1964), who suggested that infragravity waves consist of a mixture of bound and free long waves, in which the bound long waves associated with incident wave groups are somehow released as free long waves due to wave breaking and then reflected from the beach. An alternative mechanism was presented by Symonds et al. (1982), who proposed that temporal variations of breaking point induced by incident wave groups may result in oscillations of wave setup, producing free infragravity waves. Nonlinear interactions between pairs of incident wave components during shoaling would also excite forced secondary waves with infragravity periods (Herbers and Burton, 1997; Herbers and Guza, 1994; Madsen et al., 1997). Overall, it is a consensus that the generation mechanisms of infragravity waves are strongly correlated with nonlinear wave shoaling and breaking processes of incident wave groups.

Few laboratory experiments and field measurements have been conducted to investigate infragravity wave processes on the reefs. For example, Nwogu and Demirbilek (2010) carried out a series of

* Corresponding author. Tel.: +886 226215656x2098.
E-mail address: ssf@mail.tku.edu.tw (S.-F. Su).

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laboratory measurements and found that the infragravity wave energy at the reef flat increases with increasing incident wave energy and rising tidal water level. They showed that the infragravity wave energy increases shoreward from the reef edge to the shoreline. Péquignet et al. (2009) also observed from field measurements that the largest infragravity energy is located close to the beach at the high tidal level during the tropical storm event.

Numerical models have been employed to study wave transformation and infragravity wave processes on steep fringing reefs (Filipot and Cheung, 2012; Nwogu and Demirbilek, 2010; Roeber and Cheung, 2012; Sheremet et al., 2011; Yao et al., 2012; Ma et al., 2014). Filipot and Cheung (2012) applied the phase-averaged wave model SWAN (Booij et al., 1999) to simulate wave transformation over the fringing reefs. With the optimal parameterizations of wave-breaking and bottom-friction for coral reefs, the model is able to accurately predict the wave height and setup, but unable to reproduce infragravity energy due to the model limitation in triad-interaction approximation. Sheremet et al. (2011) developed the phase-resolving and phase-averaged wave models based on the nonlinear mild-slope equation with a steep-slope correction to predict wave spectral transformation over fringing reefs. Validation of the models against the laboratory experiments of Demirbilek et al. (2007) indicated that both models accurately predicted swell and sea energies. However, the infragravity wave energy was underestimated, and the predicted speed of infragravity waves was slower than that in the experiments. Other phase-resolving models, such as Boussinesq equation models (Nwogu and Demirbilek, 2010; Roeber and Cheung, 2012; Yao et al., 2012) and non-hydrostatic wave models (Ma et al., 2014), have also been utilized to investigate surface wave processes and cross-spectrum energy transfer on fringing reefs. Although the capability of modeling infragravity energy has been shown in the above-mentioned models, the spatial variations and quantification of infragravity motions have not been fully investigated.

The reef bed roughness, tidal water level as well as reef geometry (e.g. the forereef slope) are the major factors affecting the evolution of infragravity waves on the reef flat. Field observations have demonstrated that the bottom friction significantly influences wave energy dissipation (Lowe et al., 2005; Pomeroy et al., 2012; Rosman and Hench, 2011). Lowe et al. (2005) isolated the relative contributions of wave breaking and frictional dissipation in a field coral reef platform using the numerical model REF/DIF 1. They found that wave energy dissipation by bottom friction is comparable to wave breaking in the study site. A numerical study by Van Dongeren et al. (2013) using the XBeach model also showed that the success of the model simulation of infragravity waves is relied on a large bed friction coefficient. The effects of tidal water level on infragravity wave motions were investigated by Pomeroy et al. (2012), who found that infragravity wave energy on the reef flat is strongly dependent on the tidal water level. The infragravity wave energy increases with rising tidal water level. Yao et al. (2012) studied the effects of forereef slope on wave transformation in the reef platform. They found that the forereef slope has a negligible effect on wave height distribution and wave setup, but may affect the location of breaking point.

Despite the above-mentioned studies, the knowledge on how these factors affect the spatial variability of infragravity waves is still limited. For example, how these factors affect the cross-spectral energy transfer is unclear. The specific objectives of the present study are to: (1) develop a numerical model that is capable of simulating wave shoaling, breaking as well as infragravity wave generation and dissipation on fringing reefs; (2) understand the effect of incident wave height and period, reef roughness, tidal water level as well as forereef slope on spatial variability of infragravity waves over fringing reefs. The two-dimensional Boussinesq wave model FUNWAVE-TVD (Shi et al., 2012) is employed in the present study, which is a fully nonlinear and dispersive wave propagation and coastal impact model, with a robust parameterization of dissipation caused by wave breaking and bottom friction, as well as a robust moving shoreline algorithm. The model simulations will be calibrated and validated against the laboratory measurements (Smith et al., 2012), which were conducted using two different reef bathymetries with smooth and rough bed surfaces at three tidal water levels.

The remainder of the paper is organized as follows. The mathematical formulation of FUNWAVE-TVD is briefly introduced in Section 2. The laboratory experiments and model setup are presented in Section 3. A description of model validation and verification, comparisons of measured and predicted wave heights, setup and infragravity waves over the fringing reefs with different reef bottom roughness, forereef slopes and tidal water levels are given in Section 4. The effects of various factors on infragravity wave distributions over the reef are discussed in Section 5. Finally, the conclusions are summarized in Section 6.

2. Model description

In this section, the fully nonlinear and dispersive wave model FUNWAVE-TVD developed by Shi et al. (2012) is briefly introduced. The model solves the fully nonlinear Boussinesq equations of Chen (2006) using a hybrid finite volume-finite difference scheme, and incorporates a moving reference level as in Kennedy et al. (2001). The major difference between FUNWAVE-TVD and the original FUNWAVE model is that the governing equations in FUNWAVE-TVD are organized in a well-balanced conservative form and numerically solved by a high-order shock-capturing TVD scheme, which has been successfully applied in a suite of coastal engineering models (Shi et al., 2012; Ma et al., 2012). This improvement allows the model to be capable of simulating wave breaking and associated wave energy dissipation without relying on empirical formulations. In addition, the model employs a third-order Strong Stability-Preserving Runge–Kutta scheme for adaptive time stepping. The moving shoreline is captured by a robust wetting-drying scheme. With these improvements, the model has been shown to be more robust in predicting wave processes in the nearshore, including wave shoaling, refraction, diffraction, breaking as well as wave run-up/rundown on the plane and natural beaches (Shi et al., 2012).

FUNWAVE-TVD solves the well-balanced conservative governing equations, which are given by

$$\eta_t + \nabla \cdot \mathbf{M} = 0 \tag{1}$$

$$\begin{aligned}\mathbf{M}_t + \nabla \cdot \left[\frac{\mathbf{MM}}{H}\right] + \nabla\left[\frac{1}{2}g(\eta^2 + 2h\eta)\right] &= H\left\{\overline{\mathbf{u}}_{2,t} + \mathbf{u}_\alpha \cdot \nabla\overline{\mathbf{u}}_2 + \overline{\mathbf{u}}_2 \cdot \nabla\mathbf{u}_\alpha - \mathbf{V}'_{1,t} - \mathbf{V}''_1 - \mathbf{V}_2 - \mathbf{V}_3 - \mathbf{R}_s - \mathbf{R}_f\right\}\\ &\quad + g\eta\nabla h\end{aligned} \tag{2}$$

where $\nabla = ((\partial/\partial x), (\partial/\partial y))$ is the horizontal gradient operator, η is the free surface, h is the water depth from the datum, $H = h + \eta$ is the total local water depth. The horizontal volume flux is

$$\mathbf{M} = H\{\mathbf{u}_\alpha + \overline{\mathbf{u}}_2\} \tag{3}$$

$\mathbf{u}_\alpha$ is the velocity at a reference elevation $z = z_\alpha$, in which $z_\alpha = \zeta h + \beta\eta$, with $\zeta = -0.53$ and $\beta = 0.47$. The depth-dependent correction of the velocity at $O(\mu^2)(\mu = h/L$, where L is the wave length) is

$$\mathbf{u}_2(z) = (z_\alpha - z)\nabla A + \tfrac{1}{2}(z_\alpha^2 - z^2)\nabla B \tag{4}$$

where$A = \nabla \cdot (h\mathbf{u}_\alpha)$ and $B = \nabla \cdot \mathbf{u}_\alpha$. $\overline{\mathbf{u}}_2$ is the depth-averaged contribution to the horizontal velocity field, which is given by

$$\overline{\mathbf{u}}_2 = \frac{1}{H}\int_{-h}^{\eta} \mathbf{u}_2(z)dz$$