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Hydrostatic versus non-hydrostatic modeling of tsunamis with implications for insular shelf and reef environments

Yefei Bai, Kwok Fai Cheung *

Department of Ocean and Resources Engineering, University of Hawaii at Manoa, Honolulu, HI 96822, USA

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ABSTRACT

Dispersion is known to separate frequency components and reduce the amplitude of tsunami waves in the open ocean. The present paper elucidates a reverse process, in which the dispersed wave components from trans-oceanic propagation of the 2011 Tohoku tsunami regroup over the insular shelves and reefs of Hawaii. The non-hydrostatic and hydrostatic modes of a nonlinear long-wave model describe the pertinent processes with and without dispersion using the same system of two-way nested grids for direct comparison. Both solutions have very similar amplitude spectra across the ocean and reproduce the dominant components of recorded surface elevations and currents in Hawaii despite their markedly different offshore waveforms. The hydrostatic solution shows early arrival of short-period signals from nonlinear scattering around the Hawaiian Islands and faster attenuation of the amplitude due to their local source. The short-period trailing waves from dispersion more effectively excite shelf and reef oscillations giving rise to better agreement with both the amplitude and phase of the recorded data. Even with larger offshore wave amplitude, a hydrostatic model trends to underestimate the surge elevation and current velocity at coastlines fronted by shallow reefs susceptible to resonance oscillations associated with edge waves.

1. Introduction

Nonlinear shallow-water models have been the primary tool for tsunami hazard assessment and warning guidance products development (Shuto and Goto, 1978; Liu et al., 1995; Titov and Synolakis, 1998). These explicit numerical schemes provide efficient solutions for large computational problems enabling their widespread application among engineers and scientists. The hydrostatic formulation, however, does not include vertical flow nor wave dispersion. Although tsunamis are primarily shallow-water waves, observations have shown effects of dispersion characterized by separation of frequency components and development of short-period trailing waves during trans-oceanic propagation (Hanson and Bowman, 2005; Saito et al., 2010; Song et al., 2012). Comparison of Boussinesq and nonlinear shallow-water model results shows omission of dispersion leads to closer alignment of the frequency components with the leading wave and overestimation of the amplitude in the open ocean (Grilli et al., 2007; Kirby et al., 2012; Saito et al., 2014).

The effects of dispersed tsunami waves at the coast are less well studied and the findings are not as conclusive (Horrillo et al., 2006; Ioualalen et al., 2007; Kilinc et al., 2009). Some relate the overestimation of offshore wave amplitude to more severe coastal impact, supporting the use of nonlinear shallow-water models to provide conservative predictions for practical application. However, the short-period trailing waves from dispersion are spread across the ocean and convoluted with the primary frequency components through nonlinear processes over the shelf. The resulting broad-banded waves are effective in exciting multi-scale resonance with coupled oscillations from open shelves to embayments (Bellotti et al., 2012a, 2012b, Cheung et al., 2013; Yamazaki and Cheung, 2011). In addition, shallow coastal reefs in tropical environments are susceptible to edge-wave modes down to the infragravity band (Bricker et al., 2007; Roeber et al., 2010; van Dongeren et al., 2013). An acoustic Doppler Current Profilers (ADCP) detected persistent and energetic surges with periods as short as 5 min albeit with a predominant 43-min surface signal at the Honolulu coast after the 2011 Tohoku tsunamis (Yamazaki et al., 2012).

There is an increasing awareness of the need to include tsunami loads for design and retrofit of coastal infrastructure (e.g., Cheung et al., 2011; Yim et al., 2014). Understanding of dispersion on nearshore wave action provides insights into the coastal impact and facilitates proper selection of numerical models for practical application. The 2011 Tohoku tsunami, which exhibited strong dispersion characteristics across the ocean and produced comprehensive surface and current measurements in Hawaii, provides a useful case study. A continuous description of the dispersion processes from the open ocean to the coast is necessary to account for the coupled multi-scale oscillations. The nonlinear long-wave model, NEOWAVE, can describe tsunami generation, propagation, and inundation through a system of two-way nested grids without external data transfer (Yamazaki et al., 2009, 2011a). We utilize the hydrostatic and non-hydrostatic modes of

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NEOWAVE to investigate effects of dispersion on tsunami propagation in the open ocean and transformation over the inter-connected insular shelves of the Hawaiian Islands. The resonance over the shelves and reefs amplifies selected frequency components for inter-comparison of computed and recorded data and inference of dispersion effects at the coast.

2. Tsunami modeling

NEOWAVE of Yamazaki et al. (2009, 2011a) builds on the nonlinear shallow-water equations with a vertical velocity term to account for flows over steep slope and weakly-dispersive waves. The governing equations are written in spherical coordinates (ξ, ψ, z), in which ξ is longitude, ψ is latitude, and z is distance normal to the mean sea level. Let R, Ω, and g denote the earth radius, angular velocity, and gravitational acceleration; ρ the water density; and n the Manning coefficient representing the bottom roughness. The evolution of the flow in time t over varying depth d follows the continuity equation

$$\frac{\partial \zeta}{\partial t}+\frac{1}{R\cos\psi}\frac{\partial (UD)}{\partial \xi}+\frac{1}{R\cos\psi}\frac{\partial (V\cos\psi D)}{\partial \psi}=0 \tag{1}$$

and the momentum equations in the ξ, ψ and z directions

$$\frac{\partial U}{\partial t}+\frac{U}{R\cos\psi}\frac{\partial U}{\partial \xi}+\frac{V}{R}\frac{\partial U}{\partial \psi}-\left(2\Omega+\frac{U}{R\cos\psi}\right)V\sin\psi = -\frac{g}{R\cos\psi}\frac{\partial \zeta}{\partial \xi}-\frac{1}{2}\frac{1}{R\rho\cos\psi}\frac{\partial q}{\partial \xi}-\frac{1}{2}\frac{q}{D\rho R\cos\psi}\frac{\partial(\zeta-d)}{\partial \xi}- n^2\frac{g}{D^{1/3}}\frac{U\sqrt{U^2+V^2}}{D} \tag{2}$$

$$\frac{\partial V}{\partial t}+\frac{U}{R\cos\psi}\frac{\partial V}{\partial \xi}+\frac{V}{R}\frac{\partial V}{\partial \psi}+\left(2\Omega+\frac{U}{R\cos\psi}\right)U\sin\psi = -\frac{g}{R}\frac{\partial \zeta}{\partial \psi}-\frac{1}{2}\frac{1}{\rho R}\frac{\partial q}{\partial \psi}-\frac{1}{2}\frac{q}{D\rho R}\frac{\partial(\zeta-d)}{\partial \psi}- n^2\frac{g}{D^{1/3}}\frac{V\sqrt{U^2+V^2}}{D} \tag{3}$$

$$\frac{\partial W}{\partial t}=\frac{q}{\rho D} \tag{4}$$

where ζ is the surface elevation, $D = d + \zeta$ is the flow depth, and (U, V, W) and q are the depth-integrated velocity and non-hydrostatic pressure. If the non-hydrostatic pressure $q = 0$, the vertical velocity in the momentum equation (4) vanishes. The governing equations (1) to (3) reduce to the nonlinear shallow-water equations.

A semi-implicit approach evaluates the hydrostatic and non-hydrostatic components of the governing equations at each time step (Zijlema and Stelling, 2008). A staggered finite-difference scheme integrates the continuity equation (1) for ζ and the hydrostatic terms in the horizontal momentum equations (2) and (3) for U and V. The shock-capturing method of Stelling and Duinmeijer (2003) is adapted with a first-order upwind scheme to provide the advection speed for modeling of flow discontinuities. The vertical momentum equation (4) is approximated by a linear distribution of W and expressed in terms of the kinematic boundary conditions at the free surface and seabed as

$$W_{\zeta}=\frac{\partial \zeta}{\partial t}+\frac{U}{R\cos\psi}\frac{\partial \zeta}{\partial \xi}+\frac{V}{R}\frac{\partial \zeta}{\partial \psi} \tag{5}$$

$$W_{-d}=-\frac{U}{R\cos\psi}\frac{\partial d}{\partial \xi}-\frac{V}{R}\frac{\partial d}{\partial \psi} \tag{6}$$

The non-hydrostatic pressure q is determined from the vertical momentum equation (4) via a Poisson equation. This allows update of (U, V) from integration of the non-hydrostatic terms in the momentum equations (2) and (3) and ζ from the continuity equation (1) before advancing to the next time step. The terms associated with the surface and bottom slopes in Eqs. (5) and (6) are generally negligible, but may strongly influence local dispersion at shelf breaks and around large seamounts and canyons as inferred from the three-dimensional model results of Choi et al. (2007).

The modeling of the 2011 Tohoku tsunami utilizes up to five levels of two-way nested grids with increasing resolution to capture multi-scale processes from the open ocean to the shore. Fig. 1 shows the grid layout and serves as a location map for subsequent reference. The level-1 grid resolves the tsunami across the Northern Pacific at 2-arcmin (~3600 m around Hawaii), while the level-2 grid describes the transformation along the Hawaiian Islands at higher resolution of 24 arcsec (~720 m). The level-3 grids cover Kauai, Oahu, and Maui at 3 arcsec (~90 m) and Hawaii Island at 6 arcsec (~180 m). An intermediate level of grid brings the resolution of Hawaii Island's northeast-facing shore to 3 arcsec (~90 m). The finest grids at levels 4 or 5 cover between 7 and 11 km of coastlines and resolve the reefs, channels, and harbors at 0.3 arcsec (~9 m). The bathymetry and topography come from ETOPO1, multibeam, and LiDAR data at 1 arcmin (~1800 m), 50 m, and 1–3 m resolution. Hydrographic surveys provide the bathymetry inside harbors and marinas, where the water lacks clarity for LiDAR. The semi-diurnal tides have a range of 0.3 m in Hawaii and the low value justifies the use of the mean-sea level for modeling of near-shore edge waves down to the infragravity band. A Manning coefficient of 0.035 describes the subgrid roughness for the nearshore reefs and the volcanic substrates of Hawaii (Bretschneider et al., 1986).

The tsunami source is defined by a finite-fault model based on inversion of global seismic waves and iterative forward modeling of tsunami waveforms at near-field water-level stations (Yamazaki et al., 2011b). The model rupture has large slip of up to 62 m along the trench as inferred independently from geodetic data (Ito et al., 2011; Simons et al., 2011). Implementation of the planar fault model of Okada (1985) provides the time history of earth surface deformation over the rupture duration of 2.5 min. The vertical velocity term in the non-hydrostatic mode of NEOWAVE facilitates modeling of tsunami generation and transfer of kinetic energy from seafloor deformation to the ocean. For the hydrostatic solution, the vertical velocity term is disabled after the tsunami is generated and the resulting nonlinear shallow-water equations are integrated explicitly for the rest of the computation. The use of the same tsunami generation mechanism in the two solutions allows direct comparison of the results to quantify dispersion effects on trans-oceanic propagation and near-shore transformation. The time step varies from 1 to 0.05 s through the multi-level nested grid systems. The computation covers 13 h of elapsed time for development of resonance oscillations around the Hawaiian Islands. Fig. 1 also shows the locations of selected near-shore and offshore water-level stations and ADCPs around the Hawaiian Island that recorded the tsunami for evaluation of the hydrostatic and non-hydrostatic solutions.

3. Hydrostatic versus non-hydrostatic solutions

NEOWAVE is applicable to weakly-dispersive long waves due to linear approximation of the vertical velocity profile in the momentum equation (4). The capability to describe dispersion is important to development of sub- and super-harmonics through nonlinear interactions (Bai and Cheung, 2013a). Fourier analysis of the governing equations (1) to (4) with a constant water depth gives the dispersion relation

$$c^2=\frac{gd}{1+\frac{1}{4}(kd)^2} \tag{7}$$

where c is celerity, k is wave number, and kd denotes the depth