

An investigation of the morphodynamic change of reef islands under monochromatic waves

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Abstract

The persistence and habitability of coral reef islands in future extreme oceanographic conditions has received increasing attention in the recent decade, concerning that the sea level rise (SLR) and more frequent and intense storms in the context of global climate change are expected to destabilize those islands. Here, we conduct a set of wave-flume laboratory experiments focusing on the morphodynamic change of reef islands to varying ocean forcing conditions (wave height and SLR). Subsequently, a phase-resolving XBeach numerical model is adopted to simulate the monochromatic wave process and its associated sediment dynamics. The adopted model is also firstly validated by laboratory experimental results as reported in this study. It is then used to examine the impacts of island morphological factors (island width, island height, island location and island side slope) on the island migration. The combined laboratory/physical and numerical experiment outputs suggest that reef islands can accrete vertically in response to the sea level rise and the increased storminess.

Key words: coastal morphodynamics, sea level rise, reef island, wave-flume experiment, XBeach model

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1 Introduction

Coral reef islands (also known as “motu” in Polynesian language) are defined as sandy or gravel islands located on the reef platforms (Masselink et al., 2020), which are built by unconsolidated reef-derived carbonate sediments. Such features as characterized with low elevation, small area, sensitivity to the surroundings, and high population density have made them attract global concern about their future stability and existence, especially for those mid-ocean atoll nations (Kench et al., 2015; Nicholls and Cazenave, 2010; Nurse et al., 2014). The persistence and habitability of reef islands are commonly considered to be impacted by the global climate change, profoundly the long-term sea level rise (SLR) (Dickinson, 2009; Storlazzi et al., 2015, 2018) and short-term tropical cyclones (Etienne and Terry, 2012; Kench et al., 2006; Sengupta et al., 2021a, b). Projection of ongoing SLR (Kopp et al., 2014; Deconto and Pollard, 2016) and increased storm magnitude (Shope et al., 2016) are expected to increase both the magnitude and frequency of coastal flooding (Quataert et al., 2015; Storlazzi et al., 2018) and wave erosion that destabilizing reef islands (Grady et al., 2013). Figure 1 shows an artificial land-fill project at Nanhua reef in the South China Sea was destroyed by typhoon “Moli” in late December 2015. However, recent studies have shown that reef islands are capable to self-adjust its morphology in response to the changing sea level and wave condi-

tion (Costa et al., 2019; Masselink et al., 2020; Tuck et al., 2019b). Hence, most reef islands are likely to be stable or even enlarged when facing such extreme conditions, indicating that those islands can persist on the reef platforms in the future (McLean and Kench, 2015; Kench et al., 2015, 2018; Duvat, 2019). Therefore, a better understanding of the physical processes controlling the morphological response of reef islands to SLR and storms will be essential for government agencies of atoll nations to plan adaptation and risk reduction strategies.

Previous studies tackling morphodynamic change of reef islands mostly focused on field investigations (Etienne and Terry, 2012; Kench et al., 2006, 2015; Kayanne et al., 2016; Talavera et al., 2021), leaving laboratory study on the responses of reef island morphological change to SLR and/or wave conditions rare in the field. To the best of the authors’ knowledge, Tuck et al. (2019b) remains the pioneering study to perform a series of physical laboratory experiments in a wave flume. Their physical model was built with quartz sand (median grainsize 0.35 mm) and its original morphology was constructed according to a transect surveyed at Fatato Island, Funafuti Atoll (South Pacific). They measured the island’s morphodynamic responses to changing incident significant wave height and SLR, revealing reef islands move lagoonward on reef platforms in such conditions. Their results suggested the overwash process potentially serves as the mech-

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Fig. 1. Destruction of artificial landfill project at Nanhua reef in the South China Sea under the typhoon “Moli”. Satellite imagery are sourced from Google Earth™.

anism to build and keep the island freeboard above the sea level. Their physical model was later further extended, using the entire Fatao Island as the prototype and via wave-basin experiments in Tuck et al. (2019a). They found that lagoonward island recession, vertical seaward crest accretion and spit rotation are the key forms of morphological responses of the island to the changing sea level and incident wave height. They also highlighted the magnitude and rate of such morphodynamic adjustment is closely related to the increasing sea level and wave height. More recently, Tuck et al. (2021) improved the experimental design of Tuck et al., (2019b) by gradually adding a small fraction of sediment volume to the island’s shoreline, investigating the impact of storm-driven sediment supply on the island’s morphological response. Their observations show that the sediment supply notably promotes the increase of island elevation but dampens both the island migration and the subaerial volume reduction in response to SLR. All the laboratory studies emphasize the consideration of morphodynamic processes of reef islands is urgently needed into the flood assessment models so as to provide accurate estimation of future flood risks.

Numerical models provide us alternative ways to interpret wave, current and sediment processes in reef environments. Reef hydrodynamic simulations in laboratory-scale studies have been mostly addressed with Boussinesq models such as FUNWAVE (Ning et al., 2019; Su et al., 2015; Su and Ma, 2018) and COULWAVE (Yao et al., 2012, 2016), as well as non-hydrostatic models such as NHWAVE (Ma et al., 2014; Shi et al., 2018) and SWASH (Buckley et al., 2014; Liu et al., 2021). More sophisticated models based on the Navier-Stokes equations, although less computationally efficient, have also been applied when examining wave-reef interaction problems (Franklin et al., 2013; Li et al., 2019; Yao et al., 2020, 2022). However, the morphodynamic processes have not been considered in all above models.

The open-source horizontally two-dimensional (2DH) numerical model XBeach includes both a phased-averaged surf-beat method (XB-SB; Buckley et al., 2014; Quataert et al., 2015; Van Dongeren et al., 2013) and a phase-resolving nonhydrostatic approach (XB-NH; Pearson et al., 2017; Quataert et al., 2020),

have been most frequently used in previous studies to simulate the wave processes over both field and laboratory reefs. Aside from wave and current modeling, XBeach has additional modules to simulate sediment transport and bed morphological change, and it has been successfully used to study various coastal problems such as sediment transport in beach environments (Elsayed and Oumeraci, 2017; McCall et al., 2015), wave-driven erosion of sand dunes (Berard et al., 2017; De Winter et al., 2015) and morphodynamic response of barrier islands to hurricanes (Harter and Figlus, 2017; Lindemer et al., 2010). As for the issues investigated in this study, only Masselink et al. (2020) conducted numerical simulations using the horizontally one-dimensional (1DH) form of an improved XB-NH model to reproduce the wave-flume experiments of Tuck et al. (2019b). Their simulations show that reef islands comprised of gravels are morphodynamically resilient landforms, they evolve with SLR via accreting to positive freeboard while retreating lagoonward. Such island adjustment is caused by the wave overtopping process that transfer sediments from the seaward beach face to the island top, which is consistent with the laboratory observation. Subsequently, Masselink et al. (2021) used the same model to explore the role of future reef growth in morphological response of reef islands to SLR. Their results show that although reef growth significantly reduces the wave energy at the island shoreline due to wave breaking across the platform, it does not increase the ability of islands to adjust to SLR. For the same incident wave forcing, gravel islands build up higher than sand islands due to their steeper beach face gradient leading to higher runup. In general, these studies have shown the necessity to include a morphodynamic model in assessing the climate-related impacts on the reef islands.

In the present study, we report a new series of laboratory wave-flume experiments focusing on the morphodynamic change of coral reef islands to varying ocean forcing conditions (including SLR). Our experimental design differs from that of Tuck et al. (2019b) in the following aspects: (1) carbonate sand instead of quartz sand is used to build the physical island model to avoid the scaling effect for the sediment density; (2) mono-

chromatic waves rather than spectral waves are used for which the reported overtopping and overwash processes are easier to be identified; (3) the profile of reef island is idealized as trapezoid compared to the realistic double-ridged profile of Fatato Island. In this way, the results obtained in this study can be more generalized and accommodated for various reef islands. Meanwhile, we also adopt the 1DH form of XB-NH model to numerically investigate the responses of reef island morphological change to a series of island morphological features (island width, island height, island location, and island side slope) that are not fully covered by existing laboratory (Tuck et al., 2019b, 2021) and numerical studies (Masselink et al., 2020, 2021).

The rest of this paper is arranged as follows: Laboratory setup is introduced in Section 2. The governing equations and the setup of the numerical model are described in Section 3. The analyses of experimental measurements are reported in Section 4. Numerical model validation and application for which experimental data are not available are given by Section 5. Main conclusions drawn in the present study are listed in Section 6.

2 Experimental setup

Laboratory experiments are implemented in a wave flume (40 m long, 0.5 m wide and 0.8 m high) in the Hydraulic Modeling Laboratory, Changsha University of Science and Technology. The flume is installed with a piston-type wave-maker to produce the target waves at its one end and there is a slope built with porous materials to absorb wave reflection at its other end. The sizes of both reef platform and reef island physical models are designed with a 1:50 geometrical scale according to the Froude law of similarity based on the surveyed profile of Fatato Island as reported by Tuck et al. (2019b). As shown in Fig. 2a, a horizontal reef platform (8 m in cross-shore width, 0.3 m above the flume bed) is constructed followed by a 1:6 fore-reef/back-reef slope at the sea-side/leeside edge. Behind the platform, there is a lagoon. The fore-reef slope toe is set at 20.5 m downstream of the wave-maker. The whole reef model is constructed by polyvinyl chloride (PVC) plates with their widths matching the flume width. As aforementioned, a sandy reef island consists of carbonate sand is installed on the central platform with its seaside beach toe located $D = 2$ m from the reef edge. The island profile is idealized as

a trapezoid-shape profile ($W = 1$ m for its top width, $h_i = 0.05$ m for its height, and $s = 1:6$ for its seaside and leeside beach slopes) rather than the realistic double-ridged profile observed at Fatato Island. The median diameter/grainsize of the coral sand is 0.4 mm, which is comparable to 0.35 mm of quartz sand used by Tuck et al. (2019b). By assuming spherical particles, the corresponding settling velocity (calculated by the formulations proposed by Riaz et al. (2020) for carbonate sediments) is 0.050 m/s for our coral sand with a density of 2 750 kg/m³ while it is 0.078 m/s (calculated by the formulations proposed by Cheng (1997) for quartz grains) for their quartz sand with a density of 2 650 kg/m³. The effect of reef surface roughness is not considered at this stage because it is difficult to idealize the full complexity of reef roughness in the laboratory environment.

Similar to the previous laboratory experiments (Tuck et al., 2019a, b), the experiments in this study are designed to examine the morphodynamic responses of reef island to both incident wave height and SLR, which are supposed to be the direct impacts of global climate change. Therefore, a scenario with the monochromatic wave height of $H_0 = 0.1$ m, monochromatic wave period of $T = 1.5$ s and the island freeboard (positive/negative value denotes the emerged/submerged condition) of $F = 0$ m (i.e., the reef-flat still water level $h_r = 0.05$ m) is selected as the benchmark test. Based on the 1:50 model scale, the prototype wave height and island freeboard are $H_0 = 5$ m and $F = 0$ m, respectively. Meanwhile, three additional incident monochromatic wave heights ($H_0 = 0.06$ m, 0.08 m, 0.12 m) and four freeboards ($F = 0.05$ m, 0.025 m, -0.025 m, -0.05 m) are tested, corresponding to the prototype wave heights of 3.0 m, 4.0 m and 6.0 m and island freeboards of 2.5 m, 1.25 m, -1.25 m and -2.5 m. During each test, we only change H_0 or F while keeping the other one the same as that in the benchmark test. Thus totally, 8 scenarios are examined experimentally in this study. By considering the requirement that our laboratory results can be more generalized, our tested ranges of both H_0 and F are designed to be notably larger than those tested by Tuck et al. (2019b), which are only 3 m and 4 m for H_0 as well as 0.5 m and 1.0 m for F . Particularly, the island submerged conditions (i.e., $F = -0.025$ m and -0.05 m), which are not investigated by Tuck et al. (2019b), are examined in this study.

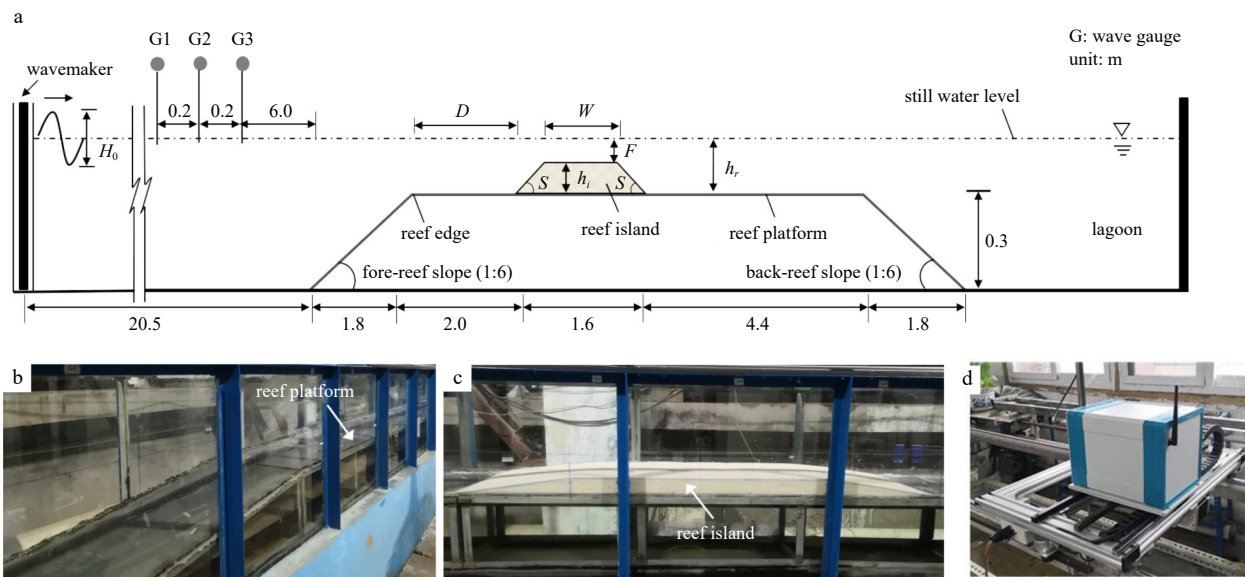


Fig. 2. Sketch of experimental setup (a); reef platform physical model (b); reef island physical model (c); laser beam profiler (d).

Three capacitance-type wave gauges (G1–G3 in Fig. 2a, Bo Ming, Ltd., Tianjin, China) are located seaward of the fore-reef slope to monitor the incident waves with a sampling frequency of 50 Hz. The reef island morphology is measured along the cross-shore central island profile by a laser beam profiler (Fig. 2d, Type LRI-III, Wuhan Tianjun Yuanda, Ltd., China). Surveys are conducted following the start of wave-maker at 1-hour interval during each test until the island reaches a morphodynamic quasi-equilibrium profile defined by less than 5% difference in the lagoonward island recession between successive surveys. A video camera (HDR-PJ675, SONY, Ltd., Tokyo, Japan) is installed on one side of the flume to record wave process around the reef island with a sampling frequency of 25 frame per second (fps). Taking the benchmark test as an example, Fig. 3a shows that the wavemaker is generally required to run for at least 6 h to achieve an equilibrium island profile. Each run is also repeated for three times, and the experimental repeatability is satisfactory in view of the whole island profile except for the rear part where a series of random sand ripples can be observed (Fig. 3b).

3 Numerical model

3.1 Governing equations

In the present study, we use the XBeach non-hydrostatic phase-resolving model (XB-NH) that can resolve various wave motions, thus it is particularly suitable for laboratory scale problem. XB-NH calculates the depth-averaged flow associated with waves and currents based on the non-linear shallow water equations. It also adds a correction into the non-hydrostatic pressure in a similar way to the SWASH model (Zijlema et al., 2011). In this study, we adopt its 1DH equations as follows, although its 2DH form is also possible,

$$\frac{\partial \eta}{\partial t} + \frac{\partial uh}{\partial x} = 0, \quad (1)$$

$$\frac{\partial u}{\partial t} + u \frac{\partial u}{\partial x} - \nu_h \frac{\partial^2 u}{\partial x^2} = -\frac{1}{\rho} \frac{\partial (\bar{q} + \rho g \eta)}{\partial x} - c_f \frac{u|u|}{h}, \quad (2)$$

where x and t denote the horizontal spatial and temporal coordinates, respectively, η is the free surface elevation, u is the depth-averaged cross-shore velocity, h is the local water depth, ρ

is the water density, ν_h is the horizontal viscosity following Smagorinsky (1963), $\bar{q}$ is the depth-averaged dynamic (nonhydrostatic) pressure scaled by the density and c_f is the bed friction coefficient. In this study, c_f is obtained based on the Manning friction law where n is the Manning roughness coefficient,

$$c_f = \frac{gn^2}{h^{1/3}}. \quad (3)$$

XB-NH includes the depth-limited wave breaking using a shock capturing scheme based on the momentum conservation. However, as a depth-integrated model, it cannot explicitly compute the overturning of plunging breakers. To determine both the magnitude and the location of wave breaking, a hydrostatic front approximation is employed where the pressure distribution underneath the breaking waves is assumed to be hydrostatic (Smit et al., 2010). There is an interval between a maximum steepness value (by default = 0.40) and a secondary steepness value (by default it is half of the specified maximum value) for the local surface steepness where the nonhydrostatic pressure correction is switched off, see Smit et al. (2013) for details.

Different sediment transport equations are adopted in the latest version of XBeach to compute bed, suspended and total sediment loads. The sediment transport process is simulated using the depth-averaged advection-diffusion equation of Galappatti and Vreugdenhil (1985).

$$\frac{\partial hC}{\partial t} + \frac{\partial hCu^E}{\partial x} + \frac{\partial}{\partial x} \left[D_h h \frac{\partial C}{\partial x} \right] = \frac{hC_{eq} - hC}{T_s}, \quad (4)$$

where C is the depth-averaged sediment concentration varying on the time scale of wave group, C_{eq} is the equilibrium sediment concentration, D_h is the horizontal sediment diffusion factor, u^E is the Eulerian wave-driven flow velocity (computed from the hydrodynamic module) in the cross-shore direction, w_s is the sediment fall velocity and T_s is an adaptation time (the elapsed time for C to reach C_{eq}) representing the sediment entrainment, which is given by a simple approximation based on h and w_s

$$T_s = \max \left(f_{T_s} \cdot \frac{h}{w_s}, T_{s,min} \right), \quad (5)$$

where f_{T_s} is a correction and calibration factor (its default value is 0.1) to consider the fact that w_s is obtained according to the depth-averaged flow condition. w_s is calculated using the formulations of Ahrens (2000) which are derived based on a relationship suggested by Hallermeier (1981),

$$w_s = \alpha_1 \sqrt{\Delta g D_{50}} + \alpha_2 \frac{\Delta g D_{50}^2}{\nu}, \quad (6)$$

$$\alpha_1 = 1.06 \tanh(0.016 A^{0.50} \exp(-120/A)), \quad (7)$$

$$\alpha_2 = 0.055 \tanh(12 A^{-0.59} \exp(-0.000 4A)), \quad (8)$$

$$A = \Delta g D_{50}^3 / \nu^2, \quad (9)$$

A is the Archimedes buoyancy index, ν is the kinematic viscosity of water, $\Delta = (\rho_s - \rho)/\rho$ with ρ_s being the density of the sediment particle. $T_{s,min}$ in Eq. (5) is the minimum adaptation time

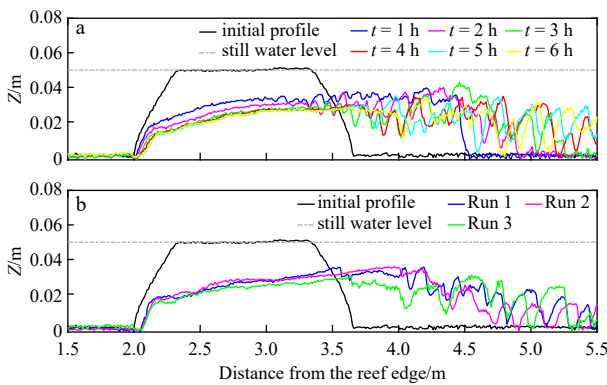


Fig. 3. Measured reef island profile every 1 h subjected to the action of monochromatic waves (a); measured equilibrium reef island profiles with three repeated runs. Z indicates the elevation relative to the reef platform surface ($H_0 = 0.1$ m, $T = 1.5$ s and $F = 0$ m) (b).

(its default value is 0.5 s). C_{eq} in Eq. (4) is the total equilibrium sediment concentration, which is calculated by

$$C_{eq} = \max \left(\min \left(C_{eq,b}, \frac{1}{2} C_{max} \right) + \min \left(C_{eq,s}, \frac{1}{2} C_{max} \right), 0 \right), \quad (10)$$

where C_{max} is a maximum allowed sediment concentration, which is a user-defined parameter (its default value is $0.1 \text{ m}^3/\text{m}^3$). The bed concentration ($C_{eq,b}$) and suspended concentration ($C_{eq,s}$) in Eq. (10) are calculated based on the Van Thiel -Van Rijn equation (Van Thiel de Vries, 2009).

The finite difference method with the first-order upwind scheme is adopted to resolve the depth-averaged sediment concentration, C , and then the cross-shore sediment transport rate, q_x is calculated according to the depth-averaged sediment concentration and the dispersion flux

$$q_x = hCu^E + D_b h \frac{\partial C}{\partial x}. \quad (11)$$

In terms of the gradient in the sediment transport, the bed level is updated according to

$$\frac{\partial z_b}{\partial t} + \frac{f_{mor}}{(1-p)} \frac{\partial q_x}{\partial x} = 0, \quad (12)$$

where z_b is the bed elevation, p is the porosity of the bed, f_{mor} is a morphological acceleration coefficient, which can accelerate the morphological time scale with respect to the hydrodynamic time scale.

3.2 Numerical setup

Referring to Fig. 2, the numerical computational domain is established to reproduce the full setup of the experimental domain. On the offshore (left) boundary, incident wave and inflow conditions are imposed. Meanwhile, a weakly reflective-type boundary condition is set on both ends of the domain to ensure the absorption of wave reflection. The time step during the computation is estimated according to a Courant number $C_r = (C\Delta t)/\Delta x$, where C is the wave speed and Δx is the grid size. $C_r = 0.7$ is employed for the subsequent simulations, yielding the numerical stability and convergence for all tested scenarios. A grid size $\Delta x = 2 \text{ cm}$ is found to be fine enough to discretize the numerical domain in the present study, which gives a total number of cells at 2 000. Our pilot test shows that no notable differences in view of the wave processes around the island as well as the island morphological change can be observed with further reduction of the grid size.

There is a group of model parameters related to both hydrodynamic and morphodynamic modules of XBeach. We try our best to keep those parameters as default values, thus only several parameters are calibrated to achieve a best match between the numerical simulations and laboratory experiments. The calibrated model parameters are obtained with Manning frictional coefficient $n = 0.075$ (its default value is 0.01), maximum wave steepness criterium $\gamma = 0.3$ (its default value is 0.4) and our coral sand density $\rho_s = 2750 \text{ kg/m}^3$ (its default value is 2650 kg/m^3). The model validation is given by Fig. 4 which shows a comparison of the equilibrium reef island profile between the numerical simulation and the laboratory measurement based on the benchmark scenario. Good agreement can be found in view of the over-

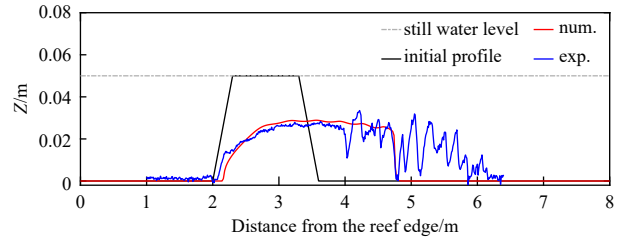


Fig. 4. Simulated and measured equilibrium reef island profiles. Z indicates the elevation relative to the reef platform surface ($H_0 = 0.1 \text{ m}$, $T = 1.5 \text{ s}$ and $F = 0 \text{ m}$).

all island profile except for the rear sand ripples. The slight mismatch around the toe of island seaside beach is mainly due to the fact that we only obtained a quasi-equilibrium profile during the laboratory experiment as mentioned in Section 2 to save the test duration.

4 Laboratory modeling results

We firstly examine measured equilibrium reef island profiles with varying incident wave height (Fig. 5). At smaller wave height when $H_0 = 0.06 \text{ m}$, a lagoonward recession of the whole island can be observed. A crest (highest point in elevation) is also built around the island's seaside beach due to the fact that wave run-up exceeds the top of island (i.e., the wave overtopping process as will be discussed later), when breaking waves are small (the incipient wave breaking has already occurred near the reef edge). As H_0 increases, the island's crest lowers significantly, thus destructive process of the island due to the overwash process can be observed under larger breaking waves (discussed later). The island crest moves to the rear part of the island and its height is reduced with the increase of H_0 . Again, a series of sand ripples can be found after the ridge associated with larger H_0 . Moreover, the seaside beach toe is found to notably migrate away from the reef edge with $H_0 = 0.12 \text{ m}$.

As for the effect of reef island freeboard due to SLR, Fig. 6 shows that the sediments near the island's seaside beach are transported to the emergent island top by wave overtopping with larger freeboards ($F = 0.05 \text{ m}$ and $F = 0.025 \text{ m}$). This leads to a crest accretion located at the central island, which is consistent with the laboratory observation of Tuck et al. (2019b), although their initial island has a double-ridged profile. Significant island-crest lowering, thus island destruction, across the reef platform occurs under the zero emergence ($F = 0 \text{ m}$), this may be the case for Nanhua reef under typhoon “Moli” as shown in Fig. 1. After that, as the island becomes more submerged, the island crest forms at the island's leeside beach. The crest moves further seaward but its elevation increases significantly as F changes from -0.025 m to -0.05 m as a result of the reduced depth-lim-

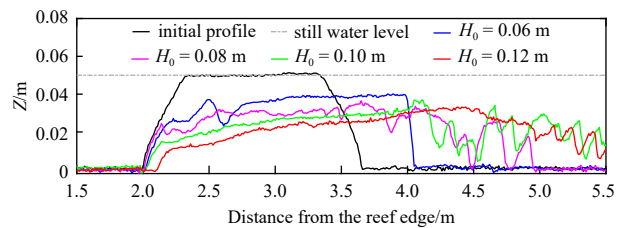


Fig. 5. Measured equilibrium reef island profiles with varying incident wave heights (H_0). Z indicates the elevation relative to the reef platform surface ($T = 1.5 \text{ s}$ and $F = 0 \text{ m}$).

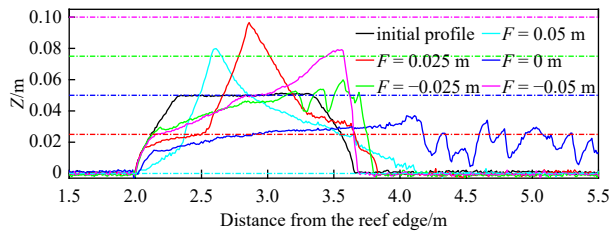


Fig. 6. Measured equilibrium reef island profiles with varying island freeboards (F). Dotted line with the same color as F indicates the corresponding still water level due to sea level rise. Z indicates the elevation relative to the reef platform surface ($H_0 = 0.1$ m and $T = 1.5$ s).

ited breaking wave process. The recession of island beach on the leeside also decreases with the enhancing submergence. Overall, our results present evidence that the island top in both the emerged and submerged conditions is able to accrete vertically in response to the rising sea level, the former is conformed to the previous laboratory experiments (Tuck et al., 2019b) and numerical simulations (Masselink et al., 2020), while the latter is our new finding. Both are expected to afford the reef island's potential to offset future sea level rise and storm event.

Previous laboratory studies with similar design (Tuck et al., 2019a, b) have identified that overtopping and overwash processes can physically account for such island evolution as can be observed in both Figs 5 and 6. This can be further verified in our experiments based on the records from the video camera. Notably, under small wave height and SLR, the overtopping process dominates, leading to an increase in the island crest elevation

(e.g., Fig. 7a). On the contrary, under large wave height and intermediate SLR (i.e., zero freeboard), the overwash process dominates, resulting in rollover of the entire island (e.g., Fig. 7b). A further increase in SLR reduces the depth-limited breaking-wave action thus weaken the overwash process.

5 Numerical modeling results

Since the effects of wave event (indicated by the wave height) and SLR on the island change have been investigated in the laboratory experiments in this study, in this section, we focus on the impacts of varying morphologic features of the island on the island migration. Four related parameters (island top width W , island height h_i , distance between island toe and reef edge D and island side slope s) are examined via numerical simulations. Ranges of tested values are summarized in Table 1. The default scenario is selected as $W = 1$ m, $h_i = 0.05$ m and $D = 2$ m. In each run, the simulation completes when an equilibrium profile is achieved. We only change the value of one parameter at one time while fixing other parameters as the defaults. The aforementioned benchmark wave condition ($H_0 = 0.1$ m and $T = 1.5$ s) and reef-flat still water level ($h_r = 0.05$ m) is used all through the simulations, which gives the freeboard $F = 0$ for all the runs except for those with varying h_i .

The island is found more difficult to be eroded responding to the increase of island's top width due to increased sediment mass as well as larger wave dissipation occurring over wider island top (Fig. 8). Thus both the lagoonward recession of the island as well as its crest lowering are weakened, indicating that the narrow island is more vulnerable to the overwash process.

As the island height increases from 0.025 m to 0.1 m (Fig. 9), the island top changes from submerged to emerged conditions. A

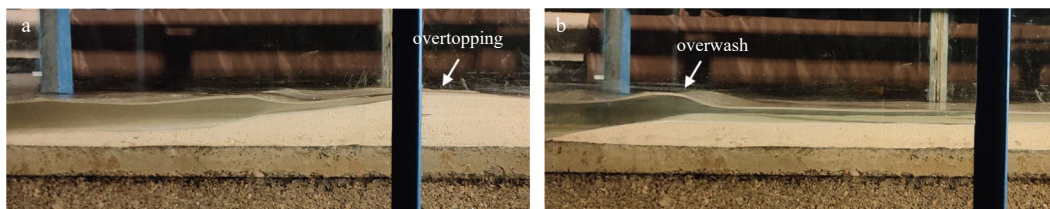


Fig. 7. Snapshots of the wave processes around the reef island during the experiments: overtopping under $F = 0.05$ m (a); overwash under $F = 0$ ($H_0 = 0.1$ m and $T = 1.5$ s) (b).

Table 1. Summarize of tested values of reef island morphological factors

Factors	Tested values
Island top width W /m	0.5, 1.0 , 1.5, 2.0
Island height h_i /m	0.025, 0.05 , 0.075, 0.1
Distance between island toe and reef edge D /m	1, 2 , 3, 4
Island side slope s	1:3, 1:6 , 1:9, 1:12

Note: Boldface indicates the defaults.

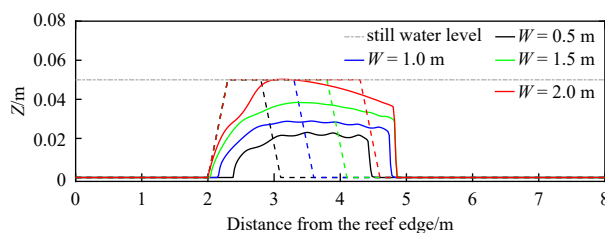


Fig. 8. Simulated equilibrium reef island profiles with varying island top widths (W). Dashed line with the same color as W indicates the corresponding initial island profile.

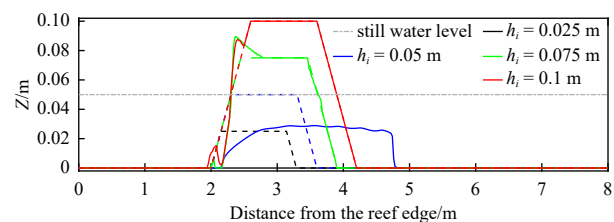


Fig. 9. Simulated equilibrium reef island profiles with varying island heights (h_i). Dashed line with the same color as h_i indicates the corresponding initial island profile.

transition from the buildup of island seaside beach (similar to scenario with $F > 0$ in Fig. 6 due to the wave overtopping) to the destruction of whole island profile (similar to $F < 0$ in Fig. 6 due to the wave overtopping) can be observed. Particularly, at the small $h_i = 0.025$ m, the island is entirely washed away.

As for the effect of the island location indicated by the distance between island beach toe and reef edge (Fig. 10), the influence of overwash process will decline as a result of enhanced energy dissipation from both wave breaking and bottom friction when travelling longer distances over reef flat. Consequently, the lagoonward island movement decreases but its crest elevation reduction is almost unaltered.

Comparatively speaking, the island morphodynamic change is less sensitive to the variation of island beach slope (Fig. 11). Both the inland recession of the island as well as the crest lower-

ing slightly decreases as the beach slope becomes steeper. This is mainly due to fact that steeper slope reflects more incident wave energy thus less energy being kept for the wave overwash process.

The variations of island migration in terms of its lagoonward recession (indicated by the difference in the seaward beach toe location between the initial profile and the equilibrium profile, ΔL) and its crest lowering (indicated by the difference in the crest elevation between the initial profile and the equilibrium profile, ΔZ_c) with the examined island morphological factors are further demonstrated in Figs 12 and 13. Note that for the scenario with island washed away, i.e., when $h_i = 0.025$ m, we have $\Delta L = \infty$, which is not shown in Fig. 12b and $\Delta Z_c = h_i$ which is shown in Fig. 13b.

In accordance with those can be observed from Figs 8–11, Fig. 12 indicates again that ΔL decreases significantly with the in-

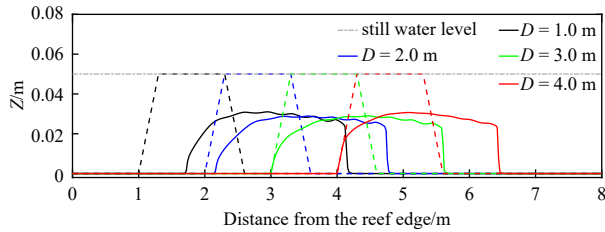


Fig. 10. Simulated equilibrium reef island profiles with varying distances between island toe and reef edge (D). Dashed line with the same color as D indicates the corresponding initial island profile.

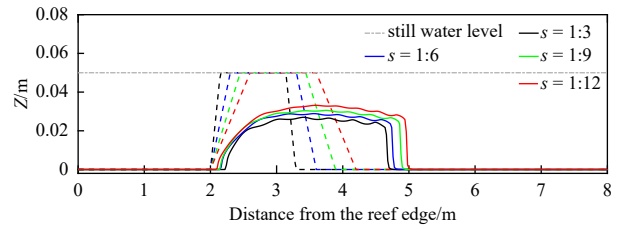


Fig. 11. Simulated equilibrium reef island profiles with varying island side slopes (s). Dashed line with the same color as s indicates the corresponding initial island profile.

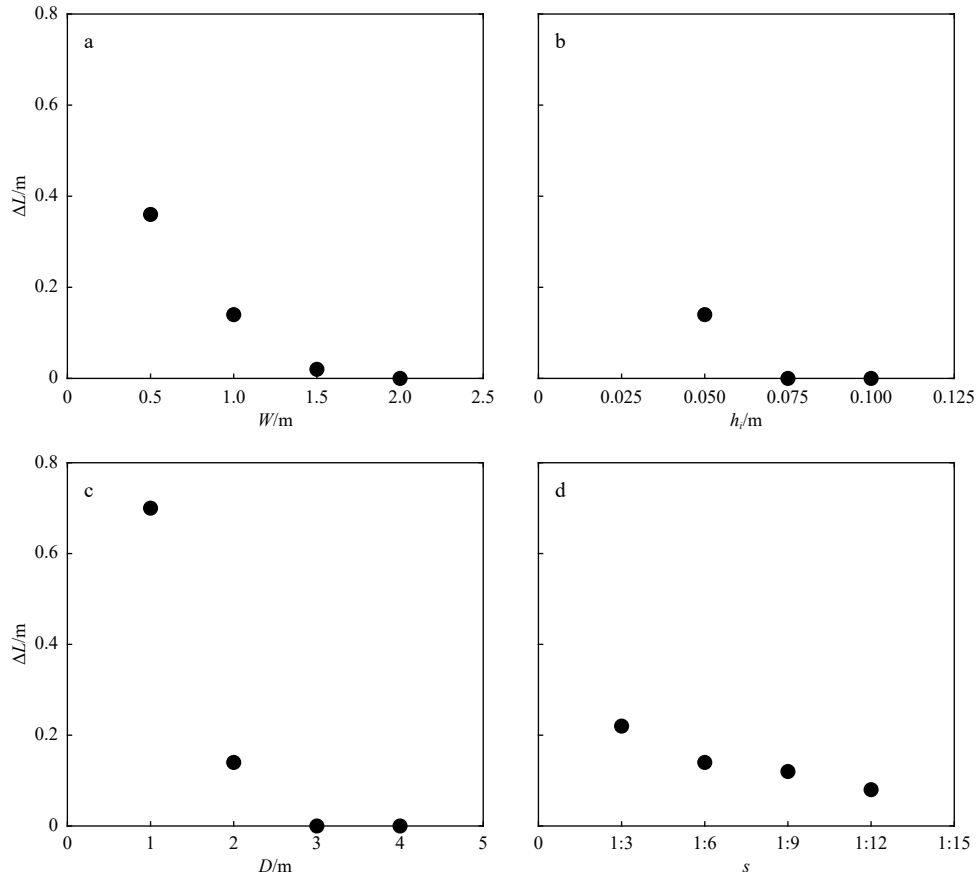


Fig. 12. The simulated lagoonward recession of seaside beach toe (ΔL) with different: island top widths (W) (a), island heights (h_i) (b), distances between island toe and reef edge (D) (c), island side slopes (s) (d).

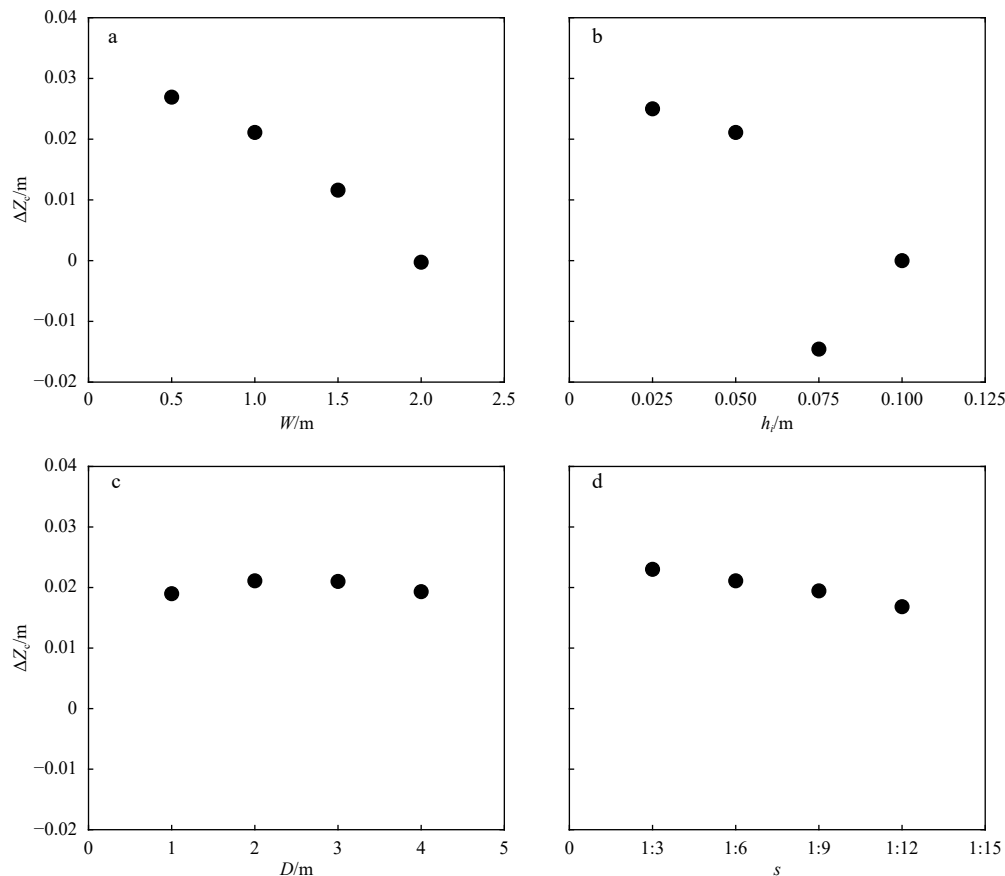


Fig. 13. The simulated island crest lowering (ΔZ_c) with different: island top widths (W) (a), island heights (h_i) (b), distances between island toe and reef edge (D) (c), island side slopes (s) (d).

crease of island width, island height, distance between island and reef edge as well as beach slope. Meanwhile, Fig. 13 shows that ΔZ_c decreases rapidly with the increase of island width but slowly with the beach slope. As for the variation of ΔZ_c with island height, a breakpoint in ΔZ_c variation can be found around the zero submergence (i.e., $h_i = 0.05$ m and $F = 0$) due to the aforementioned transition from the overwash process to the overtopping process. Comparatively speaking, ΔZ_c is not very sensitive to the variation of distance between island and reef edge within the tested range.

6 Conclusions

To better interpret wave and sediment processes over a reef island located on the reef flat subjected to the monochromatic waves, a new set of laboratory experiments is implemented in a wave flume to explore the morphodynamic change of reef islands to varying ocean forcing conditions. It is found that under small wave height and SLR conditions, the dominating overtopping process leads to an increase in the island crest elevation. Under large wave height and intermediate SLR conditions (i.e., around zero freeboard), the dominating washover process can move or even destroy the entire island. A further increase in SLR reduces the breaking-induced washover process thus a buildup of the island crest can again be found. Subsequently, the XBeach numerical model using its phase-resolving nonhydrostatic module (XB-NH) combined with its sediment transport module is adopted. The adopted model is firstly calibrated using our experimental measurement of the equilibrium reef island profile exposed to monochromatic waves. It is then used to examine the

impacts of island morphological factors on the island evolution. Analyses indicate that the reef island with smaller width, smaller height, smaller distance from the reef edge and milder beach slope is more vulnerable to the wave overwash process in view of its increased lagoonward recession and crest lowering.

Our findings in this study also show that coral reef islands can accrete vertically in response to the sea level rise and the increased storminess. Such natural adaptation may support future persistence and habitability of those reef islands, which provides valuable guidance for coastal defense purpose. Further research can be directed to incorporate realistic island profile, carbonate sediment supply, more wave regimes as well as the variation of sediment types among different islands considering their local species of corals.

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